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ARKANSAS GEOLOGICAL COMMISSION  
Norman F. Williams, State Geologist

SYMPOSIUM ON THE GEOLOGY OF THE  
OUACHITA MOUNTAINS

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VOLUME II

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Economic Geology, Mineralogy, and  
Miscellaneous

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A special debt of gratitude is expressed to Mrs. Loretta S. Chase, Donna Rinke, and L. P. Kelone for their unstinting technical assistance during the preparation of this volume.

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# F O R E W O R D

By Thomas A. Hendricks<sup>1</sup>

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

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

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

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

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

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

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

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

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

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

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PART I  
DEDICATION OF SYMPOSIUM VOLUMES  
TO THE MEMORY  
OF



**HUGH DINSMORE MISER  
(1884-1969)**

Dr. Miser frequently returned to Arkansas in his later years from his work in Washington, D. C. for consultation with members of the Arkansas Geological Commission. During his visit in November, 1968 he also met with members of THE OZARK SOCIETY, notably Joe Marsh Clark and Wellborn Jack, Jr., and related to them one of his

early journeys in the Cossatot River Country. A copy of the text was subsequently published in the Spring 1969, Volume III Number 2 of the Ozark Society Bulletin. We graciously acknowledge Joe Marsh Clark and other members of THE OZARK SOCIETY for permission to print this report in the Ouachita Symposium.

## AN 18-DAY FOOT JOURNEY IN THE HEADWATER

### COUNTRY OF COSSATOT RIVER, 1916

By Hugh D. Miser<sup>1</sup>

A foot-journey of 18 days through the scenic headwater country of the Cossatot River in July and August 1916 has proved to be an important never-to-be-forgotten milestone in my travels. This pine-forest country had not at that time seen good roads and autos, but it was inhabited by sturdy frontier people, largely homesteaders, who were uniformly kind and hospitable and gave me food and shelter.

The numerous localities I visited on the trip were in trailless forests on rugged mountains and they were accessible only on foot. My observations at these localities were a continuation of many years of geologic studies that were sponsored by the U. S. Geological Survey beginning in 1907, under the leadership of Professor A. H. Purdue of the the University of Arkansas. In 1907-08 the studies were in cooperation with the Arkansas Geological Survey. Before 1916 the geologists lived in tent camps and supplemented foot travel by riding horses and mules.

The foot trip afforded a wide variety of experiences. In the past half century those experiences have been described again and again to relatives and friends who seemed incredulous and seemed to accuse me of embellishing stories. The farther my audiences of friends were away from Arkansas the deeper was their incredulity.

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<sup>1</sup>Published by permission of *The Ozark Society* from the Spring 1969, Volume III Number 2 of *The Ozark Society Bulletin*.

The starting point was Standing Rock on Board Camp Creek where I had been brought in a buckboard from Mena. Thence I proceeded in a southerly direction across Eagle Mountain and past Shady post office to Hanna Mountain which I climbed. I descended to the east base of the mountain after sundown, crossed the Cossatot River, and reached a house at dark. I hoped to spend the night there. My anticipation seemed, for a moment, turned to disappointment when the occupant told me he was a hermit and had no accommodations for a guest. He added, "You will need to go three miles up the Cossatot River to Shady if you wish to stay at a place where there are women." My reply was, "I am not looking for women folks; I wish only to have a place to sleep." He said, "I will give you my best."

For supper he cooked bacon on a wood fire in the frontroom fireplace and served it and bakery bread on a side table in the kitchen. The table had a newspaper cover whose long use made the printed matter unreadable.

After supper he offered me his favorite rocking chair--crudely made from a flour barrel; the barrel was fastened atop two rockers; the staves were sawed in two on the front side of the chair; and the barrel's bottom was filled with quilts to form a cushioned seat. The host's conversation soon turned to an account of his hunt for wild game that day; he had shot a wolf and recovered its scalp for which he would collect a bounty from Polk County, to purchase by mail order a package of Hayners liquor from Kansas City. The Winchester that killed the wolf was



Photograph by Neil Compton

The Hanna Range — Cossatot Mountains

standing in one corner of the front room; a shot gun stood in another corner; when the host adjusted the cot on the front porch for me to sleep on he removed therefrom the longest pistol I ever saw. This arsenal aroused my apprehension which remained with me until 6:00 a.m. the next morning when we departed from his house. I walked alongside the horse he was riding. He brought with him the wolf scalp of yesterday's hunt. We traveled together up the Cossatot River past Shady post office whence he carried mail on a semi-weekly star route between Shady and Mena. Before we separated at the mouth of Mine Creek a distressed elderly lady came to the roadside to greet the mail carrier and hopefully tell him to bring her a letter from her seriously ill son in Kansas City. After we parted at noon I ate for lunch the last of my meager supply of cheese and crackers; I had become hungry and weak. That night and the next three nights George Whisenhunt and his wife, a wonderful housekeeper, gave me shelter and fine food. The day I left the Whisenhunt home I went in a southeasterly direction across uninhabited mountainous country. I had been told that black bear had come to that vicinity to feast on a bumper crop of huckleberries. I saw no bear, but I did see and enjoy eating huckleberries; some of the berries had dried and were most delicious. They reminded me of raisins.

I reached a lone house on Saline River as darkness fell. The people there could not keep me because one of the family was ill with "slow fever." I was directed to an inhabited house three miles to the south. The denseness of the forest made the road invisible through the forest. The only way I knew I was following the road was from the feel and sound of my boot hobnails striking the rocky roadbed. The road seemed to end at a cotton patch in which there was a cabin. The cabin was vacant, but I planned to spend the night there. Soon I found a foot-wide plank to use for a bed. It was placed atop a high pile of fence rails on the cabin porch and was then covered with pine boughs to soften the plank's hardness. At daybreak I dismounted from the rail pile and instantly saw a blue smoke column curling skyward from a house at the far edge of the cotton patch. An elderly couple in the house welcomed me for breakfast, they had already eaten. The gracious feeble housewife soon prepared a generous breakfast.

Most of the day was spent on Raspberry Mountain at the head waters of the Saline River. For hours near midday a heavy rainstorm accompanied by terrific lightning, thunder, and echoes from all directions enveloped and alternately darkened and lighted the mountain. I managed as best as I could to keep dry my field maps and notes. On another occasion in this region I observed how lightning had struck a tree on a mountain summit, had followed down a tree trunk and roots, and had blasted the bedrock novaculite to open a trench on whose bottom the shredded tree roots lay bare.

When I reached the valley of Blaylock Creek at the north base of Raspberry Mountain, I was still soaked to the skin from the cold rain, but my fired-up interest in this wonderful country had banished discomfort. Some years earlier a

frontier family was living in a house on a cleared tract in this valley. Now, the family was gone, so were the house and the fences, but a peach tree remained and was filled with ripe peaches. The previous day I ate dried huckleberries for lunch; today I feasted on peaches; I carried away about a dozen peaches, the maximum number I could hold in an improvised bag. Mr. and Mrs. Albert Lowry, well-known friends, of Greasy Cove on the Little Missouri River welcomed me on my arrival there. Next day Mr. Lowry rode a mule to Blaylock Creek to gather and bring home a big bag of peaches. A highlight of my 5-day stay in the Lowry home was the serving of the finest peach cobbler pies.

The Little Missouri, like the Cossatot, Saline and Caddo Rivers, is so clear that the depth of water at road crossings can not be prejudged. This was illustrated well by an experience in 1934 when Dr. Charles E. Decker of the University of Oklahoma accompanied by John Fitts of Ada, Oklahoma, and by me attempted to drive his car across a ford on the Little Missouri. The car suddenly stopped in mid-stream in knee-deep water. While Dr. Decker and I walked three miles to a CCC camp for towing service John Fitts sat in the stranded car and was entertained by watching fish schools swim around and through the hood and wheels spokes.

I left the Lowry's on a Sunday morning to visit a manganese prospect in mountains east of the Little Missouri. I needed a guide to take me to the prospect and fortunately it was Sunday so one was soon found -- a barefoot young man, three miles away from his shoes at home at Langley. He wished for his shoes to make the trip on the mountain slopes that were covered with novaculite debris having razorsharp edges. However, after brief meditation he resolved to accompany me without shoes. Amazingly, he quickly and easily climbed over the mountain slopes; he outdistanced me with my heavy hobnail boots. Never have I heard of thicker and tougher soles on human feet except at early-day iron forges in West Virginia where barefoot workers were said not to be aware they were walking on molten iron until they smelled scorched soles.

That night an elderly couple gave me shelter and food. Their cabin seemed to be so tightly and snugly built with closed windows and doors and furnished with mountain-like feather beds that I obtained their reluctant consent for me to sleep on a pallet on the front porch. The next night I spent with another elderly couple and slept in a bedroom whose windows and doors were open. The husband described his afflicted frail thin wife as having "dry consumption."

The next night I was a guest at a little hotel in Glenwood. Thence I started west on my return through Fancy Hill toward Mena.

Through advance inquiry I learned the name of a family in Fancy Hill that might keep me overnight. Luckily those good folks kept me. The next day I first went to a small

store which was said to be the community center. On my way there I met on the road two men on foot. One man was carrying a shot gun and a dead rabbit; he was recognized by me from earlier description as a man said to have escaped from the Arkansas penitentiary. After brief conversation we parted and I proceeded to the store. More than a dozen people were there. I had been forewarned that my arrival in Fancy Hill would be known promptly throughout that community. My feeling was that the people I saw on that forenoon were a kind of reception committee. After my greetings at the store were ended I made known my need for a guide to take me to a reported manganese prospect on a nearby mountain. A man kindly volunteered for that purpose. But before we left the store I purchased several watermelons, carved them with my pocket knife, and gave a slice to each person. On the way to the prospect the guide and I talked about several subjects but my companion stressed some philosophical observations to which I listened in silence. In substance he said: "For years I have worked hard to make a living but I have quit work for I have learned that I can live well without working." I clearly observed the absence of comments about the sources of funds for clothing and food for his household. We saw the prospect and then returned to the store where we parted after I expressed my sincere appreciation for his services for which I paid him.

From Fancy Hill I went in a westerly direction until approaching darkness caused me to seek and obtain shelter in a cabin on a mountainside. The kind family gave me good food for supper and breakfast. I drank coffee and no water there, for the dug water well on the hillslope below the cabin seemed to receive much of the surface drainage from the bare ground surrounding the cabin and the chicken house. As on other occasions I proposed and obtained the host's reluctant consent for me to sleep on a quilt on the front porch. I think he considered my sleeping on the porch an act of inhospitality on his part.

The next day, Saturday, was a red letter day for I wished to nearly complete my fieldwork and return on Sunday to Mena. In the forenoon I stopped at a mining camp of Edgar and Company, a prospector for manganese ore, and obtained a box lunch. About noon I stopped at a secluded spot away from the road on the Little Missouri River; I did the following things to spruce up as best as I could for my

return to Mena; I laundered my clothes (in fieldwork I carried soap for such occasions); the wet clothes were laid in sunshine on big stones to dry; a shave came next; then a bath; next I ate the box lunch; and last I dressed in dry clean clothes. Thence I went to the village of Big Fork where I stayed at a small hotel.

On Sunday forenoon I went to the vicinity of Heath School on Macks Creek to collect for Dr. E. O. Ulrich of Washington, D. C. additional graptolites like those obtained some 25 years earlier by Dr. J. Perrin Smith of the Arkansas Geological Survey. Dr. Smith's description of the graptolite occurrence indicated to me that it was on or near the road. First I looked along the road and next entered a hawthorn thicket. A barking dog chased a rabbit past me; presently a surprised local man appeared before me in the thicket. My surprise may have equaled his. We passed greetings and I identified myself. I told him I was trying to find the graptolite locality visited by Dr. Smith 25 years earlier. Promptly he led me no farther than 25 feet and remarked, "This is the slate prospect that I showed him where he might collect the fossils he wished." I recognized the kinds of fossils that Dr. Ulrich wanted me to collect. I obtained many pieces of fossiliferous slate and then improvised ways with strings and clothing to carry this additional load on my shoulders. I reached the village of Board Camp near noon and arranged by telephone for a buckboard to come there and take me to Mena some 14 miles distant.

During my wait of several hours for the buckboard to arrive I approached the store owner at his home about buying food for lunch. His response was, "I am not permitted by State law to sell you food unless you were traveling in a covered wagon." Then I indicated that a watermelon which I had seen at the store would be acceptable for lunch. The law also forbade him to sell me a melon. But he kindly indicated that I could go alone to the store, place an unspecified sum of money on the counter, and help myself to a watermelon.

The same buckboard and driver that took me 18 days earlier from Mena to Standing Rock met me at Board Camp and returned me to Mena. Thus ended a memorable experience on foot between Mena and Glenwood and through the beautiful headwater country of the Cossatot River.

P A R T I I  
T E C H N I C A L P A P E R S

GEOLOGY OF ABANDONED KELLOGG LEAD-ZINC-SILVER-COPPER MINES,  
PULASKI COUNTY, ARKANSAS

By Ronald H. Konig<sup>1</sup> and Charles G. Stone<sup>2</sup>

ABSTRACT

The abandoned Kellogg lead-zinc-silver-copper mines are located in the extreme northeastern portion of the frontal Ouachita Mountains about seven miles north of Little Rock in Pulaski County, Arkansas. These deposits were discovered in the early 1840's and mined intermittently until 1927. The mineralization occurs along a narrow 2½-mile long zone in fracture-filling quartz veins in complexly thrust-faulted sandstones and shales of the middle Jackfork Sandstone.

Three stages of mineralization are recognized: (1) the first and most important are veins containing quartz, galena, sphalerite, tetrahedrite-tennantite (occasionally freibergite variety), chalcopyrite, iron carbonate, and dickite; (2) later are veins of pyrophyllite, cookeite, iron carbonate, quartz, and possibly galena; (3) and finally numerous oxidation products.

The principal ore minerals are galena and sphalerite with lesser amounts of chalcopyrite. Assays of selected samples of galena, sphalerite, and chalcopyrite show a silver content ranging from a trace to 462 ounces per ton. A selected specimen of freibergite reportedly assayed over 1,000 ounces of silver per ton.

INTRODUCTION

The Kellogg lead-zinc-silver-copper mines, located in the extreme northeastern portion of the frontal Ouachita Mountains about seven miles north of Little Rock in Pulaski County, Arkansas (Fig. 1), were discovered in the early 1840's and mined intermittently until 1927. The prospects and shafts occur along a narrow east trending 2½-mile zone directly north of Sylvan Hills, with the principal mining activity along the northern boundary of Sec. 39, T. 3 N., R. 11 W. Examination was limited to numerous shallow prospect pits because the main shafts are either covered or flooded and, therefore, inaccessible. Samples were collected for microscopic study and analysis from these pits and nearby waste piles, and from remnants of an old ore stockpile.

Geologic quadrangle mapping by Stone (1960-1965) in the eastern frontal Ouachita Mountains included a detailed investigation of the Kellogg district. Investigations for the new State Geologic Map by Boyd R. Haley of the U. S. Geological Survey and Stone (1968-1972) further delineated the complex structure and stratigraphy of the area. The regional relationship of the Kellogg deposits to other mineral occurrences in the Ouachita Mountains of Arkansas is shown on Figure 2.

Previous Work

The first written report which briefly described the deposit was by Owen in 1858, followed by the limited

studies by Smith (1867), Conrad (1880), and Branner (1888). In 1888 Comstock described the mineralization in the district and recorded numerous chemical analyses (specifically for silver) on various samples. In 1923 Jonathan Kellogg wrote a report (unpublished) on the early history and development of the mines. The State Geologic Map in 1929 and Croneis (1930) added information on the general geology of the region. Engel (1952) described the quartz veins throughout most of the Ouachita Mountains of Arkansas. Miser and Milton (1964) and Stroud and others (1969) added further information on the mineralization in the region.

GENERAL GEOLOGY

The Kellogg mines are located within the northeastern portion of the frontal Ouachita Mountains sub-province which comprises an east-west trending structural belt about 15 miles wide, with major thrust faults, narrow commonly overturned folds, and areas of a chaotic structure. The region is bordered by moderately deformed rocks of the Arkansas Valley (Arkoma basin) on the north and intensely folded and faulted rocks of the Benton-Broken Bow uplift (core area) of the Ouachita Mountains on the south (Figs. 1 and 2).

Exposed lithic units in the region are assigned to the Paleozoic Stanley Shale, Jackfork Sandstone, Johns Valley Shale, and Atoka Formation. Tertiary rocks of the Gulf Coastal Plain province crop out about four miles east of the Kellogg mines. Granite Mountain, a Cretaceous age nepheline syenite (pulaskite) stock, is exposed south of Little Rock.

Sedimentary Rocks

The Mississippian age Stanley Shale (Table 1) usually is about 10,000 feet in thickness and typically one of

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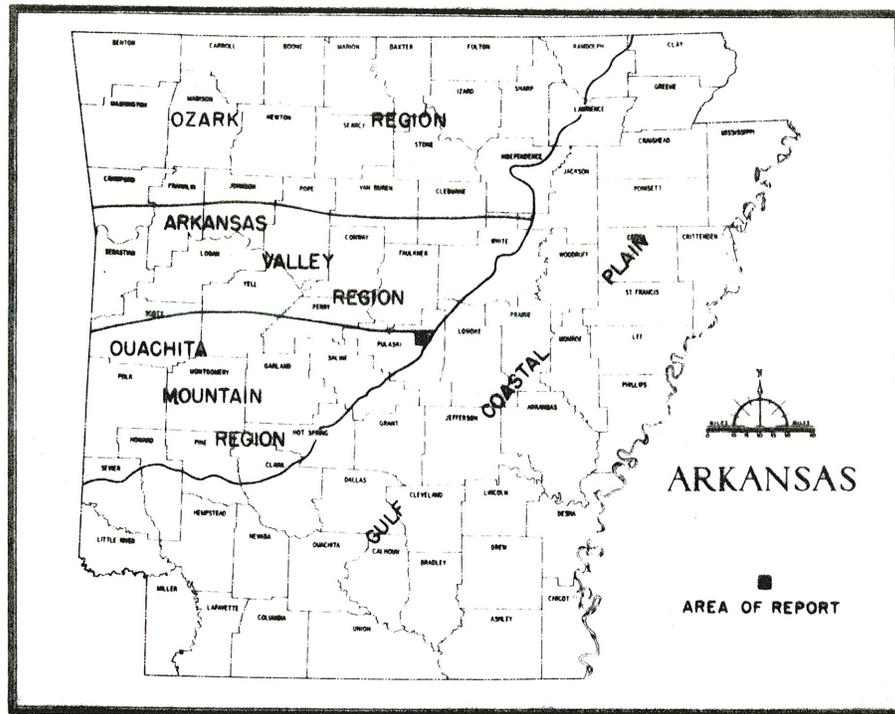


Figure 1. -- Physiographic provinces of Arkansas showing location of Kellogg mines.

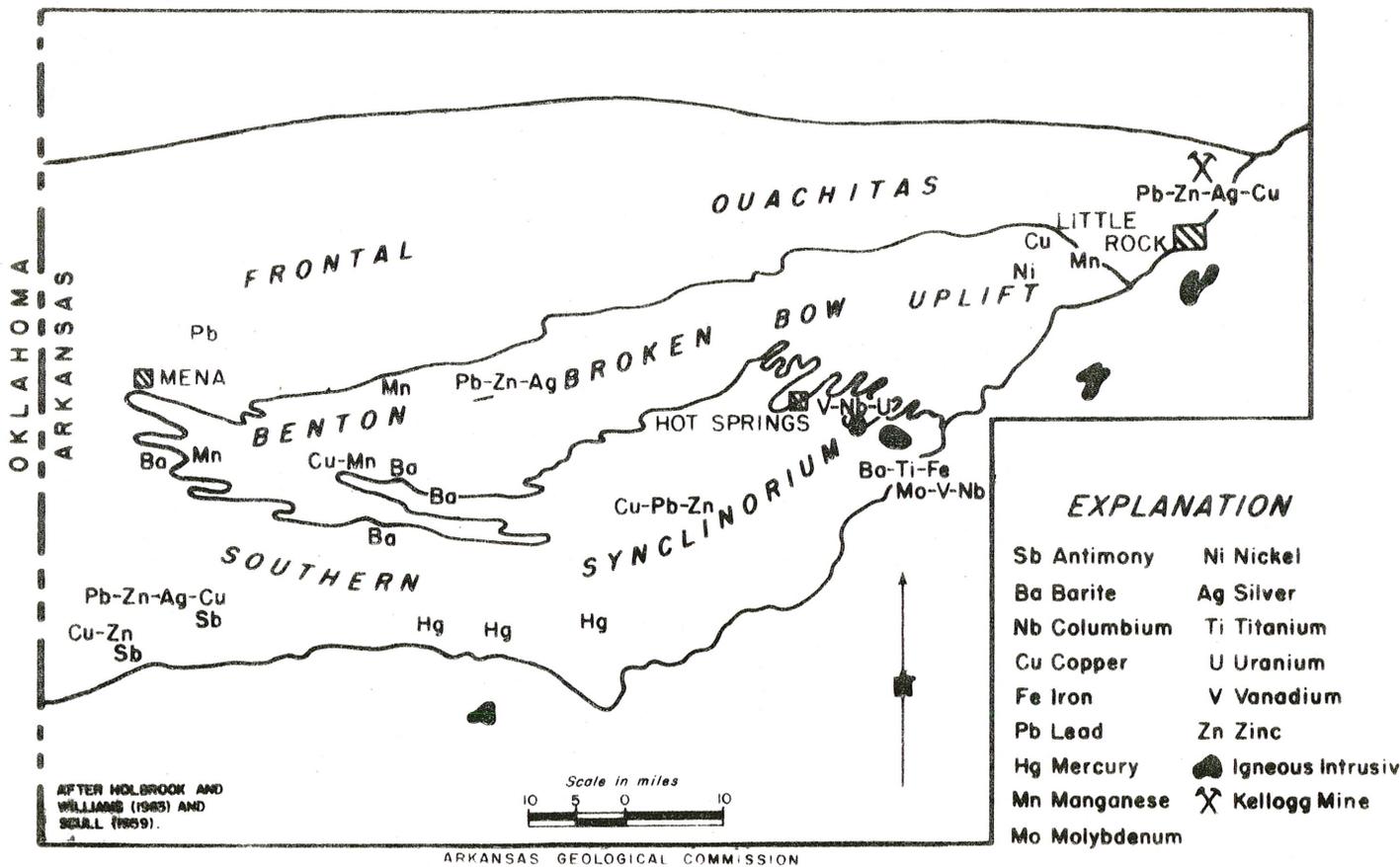


Figure 2. -- Location of mineral occurrences in the Ouachita Mountains of Arkansas.

Table 1. Stratigraphic units in the region

	System	Series	Group	Formation	Lithic Member	
<b>CARBONIFEROUS</b>		<b>Atokan</b>		<b>Atoka Formation</b>	<b>Upper</b>	
					<b>Middle</b>	
					<b>Lower</b>	
	<b>Pennsylvanian</b>	<b>Morrowan</b>			<b>Johns Valley Shale</b>	-----
					<b>Jackfork Sandstone</b>	<b>Upper</b>
						<b>Middle</b>
					<b>Lower</b>	-----
			— ? — ? — ? — ? —			
<b>Mississippian</b>	<b>Chesterian and Meramecian</b>			<b>Stanley Shale</b>		

the dominant rock units in the Ouachita Mountains, but is only partially exposed in this region because of a series of overriding thrust plates. Probably about 1000 feet of the lower Stanley crops out in isolated structures (Fig. 3) where it typically is a silty, gray-to-black shale, with some gray siliceous shale and thin subgraywacke sandstone.

The middle and upper Jackfork Sandstone forms the dominant rock exposures in the region. The lower portions of the Jackfork is apparently not exposed in the area primarily due to faulting. To the west near the Pulaski-Saline County line the lower Jackfork consists of very massive beds of light brown, quartzitic sandstone with intervals of gray shale and probably is about 1200 feet in thickness. The middle unit is probably about 3500 feet in thickness and consists mostly of silty, micaceous, reddish-gray to gray shale, but contains numerous dense, normally fine-grained, usually discontinuous intervals and masses of quartzitic sandstone. Sedimentary features indicative of turbidity current and submarine slump and slide deposition are numerous in the unit. Exposures of the middle Jackfork are intensely folded, and commonly exhibit well-developed cleavage. The Kellogg mineralization occurs in the upper portions of the middle Jackfork Sandstone.

An incomplete 2800 foot section of the upper Jackfork Sandstone is exposed south of the Kellogg mines (Fig. 3) and consists essentially of thick, alternating intervals of massive, dense, usually fine-grained, light-gray to grayish-

brown quartzitic sandstone and gray-black shale. Features indicative of turbidity current and dense sediment flow deposition occur in the lithic unit. Paleontologic data indicates that the upper portions of the middle Jackfork are Pennsylvanian (Morrowan). More scanty information on contained plant fossils (Gordon and Stone, Volume 1) suggests that the lowermost portions of the lower Jackfork are possibly Mississippian (Chesterian).

The Johns Valley Shale and the Atoka Formation overlie the Jackfork Sandstone (Table 1) in the frontal Ouachita Mountains. The Johns Valley Shale, which is about 1500 feet in thickness in western Arkansas and typically is a clay shale with small concretions, is thought to be absent in the area as a result of overriding thrust plates. The Atoka Formation is well represented north of the Kellogg mines and consists of three mappable units (lower, middle, and upper). The lower unit is exposed immediately north of the area and consists of over 13,000 feet of thin-to-occasionally massive bedded, silty, micaceous sandstone alternating with micaceous, locally sooty, gray siltstone and silty, black shale in nearly equal proportions. Features indicative of turbidity current origin typify the unit. The basal lower Atoka likely has an extreme upper Morrowan affinity.

**Igneous Rocks**

Igneous rocks are not conspicuous in the region, except for the nepheline syenite stock at Granite Mountain about

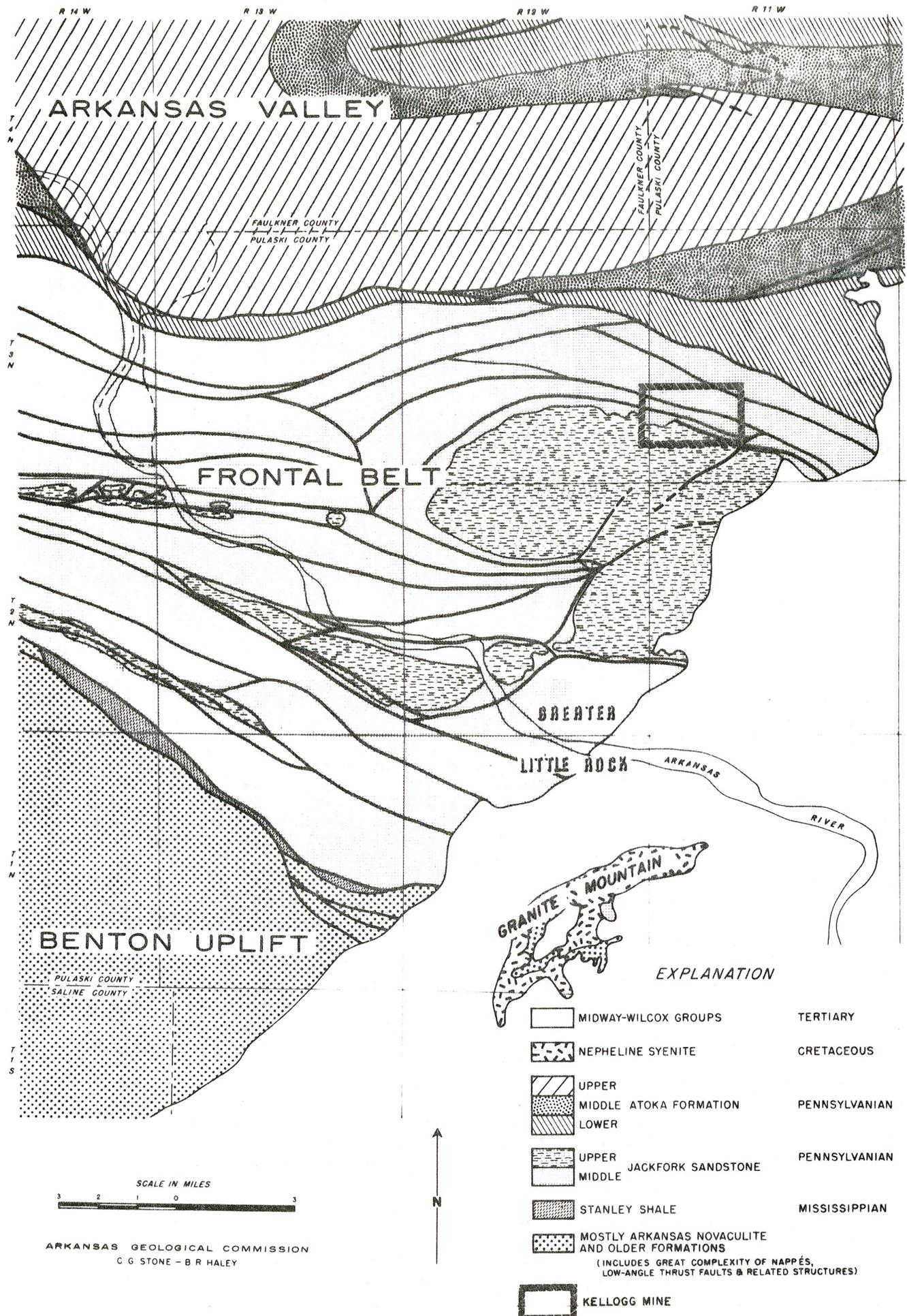


Figure 3. -- Generalized geologic map of the region around Little Rock, Arkansas.

nine miles south of the Kellogg mines (Fig. 3). Age determinations confirm this mass is an early Late Cretaceous intrusive.<sup>3</sup> Several very small, partially weathered, phonolitic-lamprophyric dikes are intruded along joints in the Jackfork Sandstone about five miles to the southwest of the Kellogg mines.

### Quartz Veins

Epithermal, fracture-filling, milky quartz veins of varying size are described by Engel (1952), Miser (1943), and others, throughout the Ouachita Mountain region and to a much lesser degree in the southern Arkansas Valley. Dickite, and less commonly, adularia are found intimately associated with the quartz. In the frontal Ouachita Mountains of Arkansas, but particularly near Little Rock, there are also small quartz veins containing needle quartz, rectorite, and cookeite - a lithium chlorite. The most notable locality is at the Jeffrey quarry in North Little Rock which was investigated by Miser and Milton (1964). Finally, some rather distinct, often smoky, quartz veins sometimes containing brookite, taeniolite and other minerals, occur in or adjacent to and are related to the Late Cretaceous igneous complexes in central Arkansas.

Most of the quartz veins in the Ouachita Mountains are essentially barren of metallic minerals, and generally reflect the composition of the enclosing wall rock, indicating that most elements of the vein minerals were acquired from the enclosing sedimentary strata. On the other hand, many of the elements comprising the metalliferous-bearing quartz veins, such as the cinnabar, stibnite, and galena in the southwestern Ouachita Mountains of Arkansas (Fig. 2) are generally considered as being derived from ascending hot solutions migrating through shear zones from unknown deeper sources.

The quartz veins containing dickite and adularia are considered by the authors as the residual products of Late Pennsylvanian - Early Permian orogenesis which culminated in the development of the Ouachita Mountains. Initially Miser and Milton (1964) thought that the cookeite-rectorite bearing veins were related to early Late Cretaceous intrusives near Little Rock; however, further studies by Stone and Milton (1976) suggest that these veins are likely Early - Middle Permian or slightly younger. Bass and Ferrara (1969) found that adularia in quartz veins at the Hamilton-Hill quartz mine in northern Garland County, Arkansas, gave isotopic ages of 214 to 287 m. y. suggesting vein deposition in Late Pennsylvanian and Early Permian time with mild terminal effects of deformation persisting into Middle Permian time or longer. Recalculations of some of these determinations by Rodger Denison

<sup>3</sup>K-Ar age determinations were made in 1963 on two concentrations of biotite from the nepheline syenite at Granite Mountain by Messrs. H. H. Thomas, R. F. Marvin, Paul Elmore, and H. Smith of the U. S. Geological Survey and gave an average age of 89 m. y. (early Late Cretaceous).

and others (see paper in Volume 1) indicate a maximum age of 263-271 million years or Early Permian.

With the foregoing determinations and data in mind, we submit that the quartz vein formation in the Ouachita Mountains likely began in very Late Pennsylvanian time and extended well into and possibly through Permian time.

### Structure

The eastern frontal Ouachita Mountains comprise an east-west trending structural belt about 15 miles wide which is characterized by major thrust faults, narrow, commonly overturned folds and areas of a chaotic structure. The Kellogg mines are situated near the northern margin of this belt and along one of a series of large thrust faults which likely displaced the Jackfork Sandstone for many miles northward over younger strata. The fault planes are poorly exposed but generally appear to dip from 45°-65° to the south and likely decrease in dip somewhat with depth. Slickensided surfaces and fractures coated with finely crystalline yellowish-white dickite are especially common in this intensely deformed belt.

Folding and faulting become progressively less intense to the north in the Arkansas Valley. Southward into the Benton uplift, Viele (1966), and Stone (1966) suggest the existence of large nappe structures. Current work by Boyd R. Haley of the U. S. Geological Survey and Stone and, in part, by Viele, indicates several epochs of complex structural deformation, including several low-angle thrust faults and the resultant windows and klippen.

All major folds and reverse faults in the region are considered to be related to the various epochs of the Ouachita orogeny, which for the most part are considered Late Pennsylvanian - Middle Permian.

### HISTORY AND PRODUCTION

The Kellogg mines were first described by Owen (1858) who states that a mining company was formed in the early 1840's but was unable to reduce the lead ore because of its silver content. The mines were reopened by the Confederates during the Civil War, in an attempt to locate lead resources. Limited information is available on these operations.

In 1888 Comstock mentions three shafts in the Kellogg area. The Welldiggen in sec. 18, T. 3 N., R. 11 W., the Confederate in SE¼ sec. 19, both of which were in bad condition due to long disuse, and the German, about 300 yards northeast of the Confederate shaft, which was being mined during his visit. The Confederate and German shafts are now thought to be located near the northern boundary of sec. 30, and the exact location of the Welldiggen is not known. The most significant development during Comstock's investigations was at the Confederate shaft, but inadequate pumping equipment caused them to avoid extensive, connected underground workings. Comstock noted that the latest exploration

was at the German Shaft where a steep slope (58°) bearing nearly south was sunk through black shales which dip about 40° south-southeast, thus cutting the beds both in dip and strike.

According to Mr. and Mrs. Harry Burton, (oral communication), caretakers of some of the mines, the main shaft (near the older Confederate and German shafts) was about 1140 feet deep with several thousand feet of tunnels and raises when all operations ceased in 1927. Partially destroyed waste dumps adjacent to the old mill site in the SE¼ SE¼ SW¼ sec. 19, T. 3 N., R. 11 W., contain about 35,000 tons of rock. Since 1927 excavations have slumped and mine and mill buildings have been largely destroyed, precluding many investigations today.

Very little information is available on the production at the Kellogg mines before 1923. It has been estimated that the Cotton enterprise (1923-1927) produced 70 short tons of lead-silver concentrates valued at about \$6000 and 30 short tons of zinc concentrates valued at \$1000. Most of this production was probably in 1925 when 3118 troy ounces of silver valued at \$2164 were obtained.

#### DESCRIPTION OF DEPOSITS

The Kellogg mines are located on the north slopes of several small hills at the southern margin of the floodplain of Kellogg Creek in portions of secs. 19, 29, and 30, T. 3 N., R. 11 W., Pulaski County, Arkansas. The known deposits occur along a linear zone which is 2½ miles long and about one-fourth mile wide. Minor traces of mineralization were also noted in small quartz veins to the west along the strike of this zone in secs. 23, 24, 25, and 26, T. 3 N., R. 12 W. In the Kellogg district there is one large shaft (NW¼ NW¼ NE¼ sec. 30, T. 3 N., R. 11 W.) with at least two smaller shafts, about 25 prospect pits and several other smaller excavations. Poor exposures in the area undoubtedly hampered the exploration of these deposits. Figure 4 is a detailed geologic map of the Kellogg district showing the location of major mines and many of the prospects.

Black, fissile, carbonaceous shale of the middle Jackfork Sandstone underlies most of the district and typically it forms slopes and valleys. Dense, quartzitic sandstone beds varying from 1 to 10 feet in thickness crop out as small ledges on the hills. Some quartz veins are also prominent along the crests of several hills. Many of the slopes are covered by a pebbly, sandy clay colluvium, which varies in thickness from a few inches up to several feet. Alluvial clays and sands 5 to 20 feet in thickness occur along Kellogg Creek.

The mineralized quartz veins occur in a highly-faulted interval composed of thin, dense quartzitic sandstone and thicker carbonaceous shale of the upper middle Jackfork Sandstone. The veins occur as fault, fracture, joint, shear or occasionally bedding-plane fillings along a general west-northwest trending zone. Cross faulting also may have

helped to control or localize some of the mineralization. The general structural pattern in the Kellogg district is shown by the cross-section in Figure 4.

Mineral zonation was possibly indicated by Comstock (1888) when he described a drift in the old German shaft 35 feet below the surface with a seam of pure crystalline sphalerite three to six inches thick. Another vein about one foot wide contains quartz and some metallic minerals. The main shaft continued 60 feet deeper to a well-defined streak of galena with minor chalcopryrite.

Most of the mineralized veins noted during the present study varied from a few inches to about four feet in width and typically pinch and swell. One vein which averaged two feet in width was traced horizontally for over 200 feet. Many of the veins appear to dip south at a fairly steep angle. Slickensides with striated surfaces coated with whitish dickite are common along many of the wall rock-vein contacts. A typical vein dips 65° S. and strikes N. 77° W. A few small veins are perpendicular to this trend.

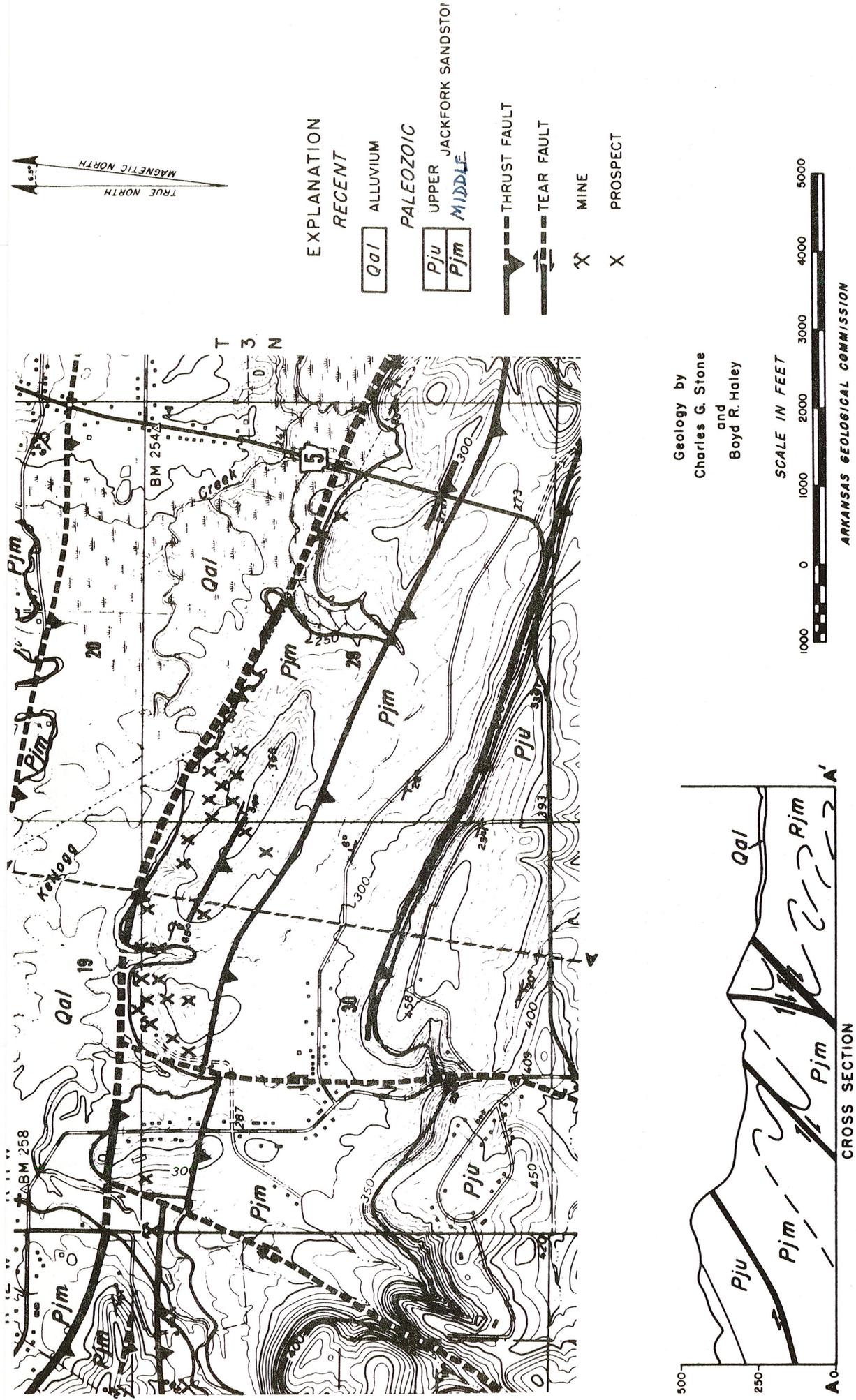
Other minor lead-zinc occurrences in the general region may be genetically related to the Kellogg deposits. Small quartz veins filling joints in the basal upper Jackfork Sandstone in the Jeffrey Quarry in sec. 8, T. 2 N., R. 12 W. (Miser and Milton, 1964) contain needle quartz crystals with abundant rectorite and cookeite, lesser quantities of ankerite, pyrite and apatite, and minor sphalerite, galena, rutile and iron chlorite. An assay of this galena did not indicate any silver.

The old McRae mine (sec. 30, T. 1 N., R. 13 W.) about seven miles southwest of Little Rock in Pulaski County, was described by Comstock (1888) as containing a peculiar fine-grained galena and pyrite in quartz. This poorly exposed occurrence is in Womble Shale of Ordovician age. One sample of galena-bearing pyritic shale contained 3.5 ounces of silver per ton.

Two probable generations of small quartz veins cutting middle Jackfork Sandstone were noted in road cuts along Ark. Hwy. 10 on the south shore of Lake Maumelle in sec. 28, T. 3 N., R. 15 W., Pulaski County. Slickensided zones contain small milky quartz veins with abundant dickite and traces of turquoise (?), siderite, and sphalerite. Small later quartz veins contain rectorite and cookeite.

#### MINERALIZATION

Examination of several hundred ore specimens from the remains of ore bins and waste piles show two stages of vein mineralization and one stage of oxidation and secondary enrichment (Table 2). The first and most important stage of quartz vein mineralization deposited dickite, quartz, iron carbonate and pyrite with the ore minerals galena, sphalerite, tetrahedrite-tennantite, and chalcopryrite. This was followed by a second stage of quartz vein mineralization resulting in deposition of pyrophyllite, cookeite, iron carbonate, quartz, and possibly galena. Finally the



Geology by  
 Charles G. Stone  
 and  
 Boyd R. Haley

SCALE IN FEET  
 0 1000 2000 3000 4000 5000

ARKANSAS GEOLOGICAL COMMISSION

- EXPLANATION
- Qal ALLUVIUM
  - Pju Pjm PALEOZOIC
  - UPPER JACKFORK SANDSTONE
  - MIDDLE
  - THURST FAULT
  - TEAR FAULT
  - X MINE
  - X PROSPECT

Figure 4. - Geologic map of the Kellogg mines area.

TABLE 2. -- PARAGENETIC SEQUENCE OF MINERALIZATION AT KELLOGG MINES

Mineral	Early Forming Minerals	Late Forming Minerals	Secondary Minerals
Pyrite	-----		Malachite Azurite
Tennantite- Tetrahedrite	-----		Cerussite Anglesite Covellite
Chalcopyrite	-----		Smithsonite
Sphalerite	-----		Linarite Brochantite
Galena	-----	----- ? -----	Hematite Limonite
Iron Carbonate	-----	-----	
Quartz	-----	-----	
Dickite	-----		
Pyrophyllite		-----	
Cookeite		-----	

primary minerals were oxidized to produce malachite, azurite, cerussite, anglesite, smithsonite, covellite, hematite, limonite, and minor linarite and brochantite.

#### Early Vein Mineralization

Minerals of this stage occur as veins and irregular masses which are largely discordant with deformed bedding in shaly horizons of the middle Jackfork Sandstone. The mineralizing solutions followed rock cleavage, shear planes, breccia zones associated with faulting, bedding, and possibly early formed joints. Deposition was mainly through filling of open spaces, although some replacement is evident in many specimens.

Dickite, the first mineral generally deposited, is always confined to either the outer extremities of veins or to the outer margins of angular fragments of wall rock contained in the veins. In both positions dickite is oriented so that (001) cleavage is either nearly parallel or perpendicular to the wall rock-vein contact.

Following dickite, milky quartz, siderite, ankerite and minor calcite developed. Iron carbonate ranges from white to pale yellowish-brown, with the darker colored varieties being siderite (sp. gr. 3.72), and the lighter colored varieties being ankerite (sp. gr. 3.17). These mineralizers also genera-

ted pyrite, chalcopyrite, sphalerite, galena, and tetrahedrite-tennantite (locally freibergite). All these minerals commonly occur in an individual vein which is separated from the country rock by a conspicuous narrow band of dickite.

Chalcopyrite (commonly iridescent) has three modes of occurrence; (1) as distinct grains scattered at random throughout the veins, (2) as narrow veins contained in tetrahedrite-tennantite, and (3) as blebs and irregular patches (exsolved) in sphalerite. Rarely these early-formed minerals occur scattered throughout restricted zones in the country rock. Microscopic examination of this material shows an abundance of angular fragments of country rock with interspersed vein material. These poorly defined zones probably represent small fault zones that have been mineralized.

The paragenetic sequence of early veins appears complex because many of these veins intersect each other even within a single polished specimen. Table 2 shows the crystallization history related to this stage of mineralization.

#### Late Hydrothermal Vein Mineralization

Later structural adjustments, resulting in the formation of fractures which intersect foliation at a high angle, are associated with renewed mineralization. This stage is

usually characterized by pyrophyllite which formed early and lines the walls of veins.

Following the formation of pyrophyllite; quartz, iron carbonate, and possibly galena developed (Table 2). These minerals occur in fractures either previously lined with pyrophyllite or in passageways by themselves. This has resulted in veins composed mainly of pyrophyllite and veins lined with pyrophyllite and containing iron carbonate, quartz and possibly galena. Only one specimen contained galena that may be of this age.

Cookeite, a lithium chlorite, has been identified in several specimens and is presumably related to this later stage of mineralization. It appears to be the last mineral formed inasmuch as it locally incrusts earlier euhedral minerals. Locally it coats fractures which intersect earlier mineralization. Cookeite and rectorite have been identified at several localities in the frontal Ouachita Mountains of Arkansas and may help to delineate this late stage of mineralization.

Approximately 95 percent of the several hundred specimens studied are related to the early stage of mineralization. This implies that the most recent mineralization was least intense and probably of lesser economic importance.

#### Oxidation and Secondary Enrichment

Secondary copper, zinc, lead, and iron in the form of malachite, azurite, covellite, cerussite, anglesite, smithsonite, brochantite, linarite, hematite, and limonitic boxwork are present in the zone of oxidation. The sulphate minerals linarite and brochantite have been identified by Charles Milton (written communication). Limonite boxwork indicates the existence of a gossan in the area, although paucity of outcrops makes it impossible to estimate the extent and degree of development of such a feature. Most of the boxwork indicates leaching of sphalerite. All secondary minerals listed above are commonly developed within the zone of oxidation during secondary enrichment. Secondary sulphide enrichment below the zone of oxidation has not been discovered, although there is no reason to assume it to be absent.

#### Silver Content of Deposits

Silver has been known in the area since the middle eighteenth hundreds. It is now established that it is contained in galena, sphalerite, chalcopryite, tetrahedrite-tennantite and freibergite. Microscopic study shows that silver does not occur as microscopically discernible minerals. Galena, however, does contain extremely thin (maximum width 0.025 mm) veins composed of microscopically undeterminable minerals. The veins are controlled largely by cleavage in galena and to a minor extent by irregular fractures. In both cases the unidentified material has formed mainly through replacement of galena. X-ray analysis of several argentiferous galena samples reveals the presence of anglesite, and possibly other unidentified minerals. It has not been determined whether these silver-

bearing compounds are confined to the veins or are scattered throughout the galena. Some silver in galena occurs as small masses of tetrahedrite-tennantite located near the join between galena and sphalerite. Locally this mineral extends from the galena into the outer margins of adjacent sphalerite. Assays of selected samples give some indication of silver concentrations; these with earlier published and unpublished assay results are summarized in Table 3. Table 4 shows some other chemical analyses made on material from the Kellogg mines.

#### ORIGIN AND AGE

The linear distribution pattern of these deposits was likely created by the ascent of mineralizing solutions along a major thrust fault zone, with the eventual localization and concentration in the highly fractured sandstone and shale, particularly, in the hanging wall.

Fracturing related to faulting, jointing, and rock cleavage is abundant in the host rock. Other physical planes of weakness appear related to bedding. Inasmuch as these planes of weakness are best developed under relatively low pressure, near surface conditions, and mineralization is largely penecontemporaneous with deformation; it is concluded that the mineral deposits also formed under similar pressure-temperature conditions. Abundant crustification, comb structure and vugs support this conclusion.

Dickite is invariably associated with the early stage of mineralization. Inasmuch as dickite has been synthesized in the laboratory under various pressures and temperatures it may be used to establish theoretical limits regarding pressure-temperature conditions. Dickite was synthesized from a  $Al_2O_3-SiO_2$  gel with water in a bomb between  $350^{\circ}C$  and  $365^{\circ}C$  by Ewell and Insley (1935). Norton (1939) formed both kaolinite and dickite under high pressure between  $250^{\circ}C$  and  $300^{\circ}C$  from feldspar and aluminum silicate minerals that were exposed to precolating water charged with  $CO_2$ . In 1954 Roy and Osborn established the upper stability temperature of  $405^{\circ}C$  for dickite by subjecting co-precipitated  $Al_2O_3-SiO_2$  gels to various temperatures and pressures (2500-25,000 lbs./in.).

Other minerals associated with the early mineralization, namely quartz, chalcopryite, galena, tetrahedrite-tennantite, siderite and ankerite are of limited value in determining the temperatures at which they may form. An exception to the above generalization involves sphalerite containing primary fluid inclusions, and blebs and patches of chalcopryite similar to that which has been observed at the Kellogg mines. Baker (1960) concluded that blebs of chalcopryite in sphalerite result from exsolution which takes place just prior to the final stages of mineralization, and may therefore be used as a geologic thermometer. Experimental work by Buerger (1934) indicates a temperature range of  $350^{\circ}$  to  $400^{\circ}C$  for the formation of sphalerite containing exsolved chalcopryite. Somewhat lower temperatures for the formation of this sphalerite have been obtained by studying bubbles contained in primary fluid

TABLE 3. -- ASSAY DATA FOR SILVER AT KELLOGG MINES

Material assayed	Silver-ounces per ton	Source of information
Assorted lead ore	30-60	Ark. Geol. Survey Rept., Vol. I
Average galena ore	27	" " " " "
Galena sample	70	" " " " "
Galena cube	\$60/ton (in 1888)	" " " " "
Galena cube	\$52/ton (in 1888)	" " " " "
Galena	462	Ark. Geol. Comm. files
Galena	11	" " " "
Galena	17	This report
Galena	17	" "
Galena	28	" "
Galena	60	" "
Chalcopyrite	10	" "
Chalcopyrite	31	" "
Chalcopyrite	7	Ark. Geol. Survey Rept., Vol. I
Freibergite	1112	" " " " "
Freibergite	788	" " " " "
Freibergite	\$1232/ton (in 1888)	" " " " "
Tetrahedrite	Trace	This report
Sphalerite	0.6	Ark. Geol. Survey Rept., Vol. I
Sphalerite	18	" " " " "
Sphalerite	18	This report
Sphalerite	0	" "
Unconcentrated ores	15.4	Ark. Geol. Comm. files
Mine run material	4.4	" " " "
Iron gossan	4.2	This report
Iron gossan	Trace	" "
Siderite	0	" "

TABLE 4. -- ADDITIONAL CHEMICAL ANALYSES FROM KELLOGG MINES

Material assayed	Element	Percent	Source of information
Galena	Au	Trace	Ark. Geol. Survey Rept., Vol. I
Chalcopyrite	Cu	26.0	" " " " "
Freibergite	Cu	5.0	" " " " "
Sphalerite-large bag	Zn	62.5	" " " " "
Unconcentrated ores	Pb	26.4	Ark. Geol. Comm. files
Mine run material	Pb	16.6	" " " "
Mine run material	Zn	5.1	" " " "
Tetrahedrite	Sb	28.2	This report
Tetrahedrite	Cu	35.0	" "
Galena	Au	not detectible	" "
Galena	Hg	0.12	" " (AA analysis)
Galena	Sb	0.32	" " " "
Sphalerite	Hg	0.10	" " " "
Sphalerite	Sb	0.17	" " " "
Siderite, trace of sphalerite	Ge	0.01	Private company (spec. analysis)
Sphalerite	Ge	0.005	" " " "
Sphalerite	Cd	0.3	" " " "
Sphalerite	Ga	Trace	" " " "

inclusions. Homogenization temperatures ranging from 168° to 230°C have been determined from a number of primary fluid inclusions. David Leach (written communication, 1974) reports a range of 160° to 228°C derived from similar inclusions in sphalerite from the Kellogg mine area, along with freezing temperatures of -4.8°C and -5.3°C for two inclusions (equivalent to 7 to 8 weight percent NaCl). This data has not been corrected for pressure because the amount of overburden present during mineralization is not known. A number of inclusions in milky quartz closely associated with sphalerite yielded homogenization temperatures ranging from 190° to 267° C.

On the basis of experimental data concerning the stability range of dickite, the temperature range in which chalcopyrite may exsolve from sphalerite, and fluid

inclusion studies, the temperature of formation of early veins at the Kellogg mines may be established. Mineralizers may have been heated to a minimum of 160°C and a maximum of 400°C, with the lower temperatures probably predominating.

Later hydrothermal mineralization, which is characterized by pyrophyllite, quartz, iron carbonate, and possibly galena, is not readily assigned a temperature of formation. Pyrophyllite has been synthesized from Al<sub>2</sub>O<sub>3</sub>-SiO<sub>2</sub> gels above 420°C under varying water pressures (Norton, 1941). However, Roy and Osborn (1954) state that pyrophyllite will also form at considerably lower temperatures with insufficient water in the system, perhaps at temperatures as low as 280°C.

A study by Bence (1964) of quartz deposits in the Ouachita Mountains of Arkansas may be used to help establish the lowest temperature of mineralization in this area. On the basis of a temperature study on liquid inclusions of selected quartz crystal specimens he found that most of the quartz in the vicinity of Little Rock crystallized at a temperature of about 225° C, a temperature that is generally higher than elsewhere in the Ouachita Mountains. Quartz specimens from the cookeite-rectorite veins filling fractures in the massive sandstones in the basal upper Jackfork Sandstone at Jeffrey quarry, five miles south of the Kellogg mines, were low, forming at a temperature of about 150°C. Inasmuch as the mineral cookeite is a conspicuous associate of both the later veins at the Kellogg mines and the veins at the Jeffrey quarry, a similar thermal history may be implied for both areas.

The major deformation of the Ouachita Mountains has been related to orogeny during Late Pennsylvanian time. Some writers have extended orogenic spasms into Permian time. The abundance of both barren or occasionally metalliferous quartz veins in this structural province has been attributed by most investigators to the genetic connection between these veins and closing orogenic spasms, with dickite and, less commonly, adularia being regarded (Miser, 1943) as an indicator of these veins. The isotopic studies of Bass and Ferrara (1969) and some recalculations by Denison and others (see paper in Volume 1) on adularia in quartz veins in the northern Ouachita Mountains of Arkansas lends strongly to the concept of quartz vein formation likely beginning in very Late Pennsylvanian and extending well into and possibly through Permian times. Cookeite-rectorite bearing quartz veins containing traces of metallic minerals are described by Miser and Milton (1964) at the Jeffrey quarry. The apparent lack of a lithium source for the cookeite, the presence of metalliferous minerals, and the apparent restriction of these veins to the vicinity of Little Rock led them to suggest that the veins probably accompanied contact metamorphism of sedimentary rocks in the vicinity of the Granite Mountain intrusives that are early Late Cretaceous in age. However, further study by Stone and Milton (1976) indicates that rectorite and/or cookeite occur in limited quantities in quartz veins westward from Little Rock along the Ouachita Mountain structural front to near the Oklahoma border, and that they do not occur in the various vein types associated with the early Late Cretaceous intrusives; an exception is the lithium mica, taeniolite, which occurs at Magnet Cove. This relationship suggests that the cookeite-rectorite veins are a product of a very Late Paleozoic heating event, possibly related to a later stage of refolding.

Most of the elements comprising the metalliferous quartz veins in the Ouachita Mountains, such as those containing cinnabar, stibnite, or galena, are considered by many investigators as Late Paleozoic in age and being derived from ascending hot solutions migrating through shear zones from unknown deeper sources. Analyses of some randomly collected Paleozoic shales from the Ouachita Mountains of Arkansas (Vine and Tourtelot,

1969) indicate trace amounts of copper, lead, and occasionally minute traces of silver, but no zinc. Stone and Milton (1976) indicate that the upper Paleozoic shales in the eastern Ouachita Mountains average about 100 ppm lithium. Thus, it would seem that the enclosing sedimentary rocks should not be completely overlooked as a possible source of many elements found in quartz veins in the region.

Igneous rocks of Paleozoic age are rare at the surface in Arkansas, being restricted to some volcanic tuff horizons in the Mississippian Stanley Shale (especially in the southern Ouachita Mountains) and to small soapstone masses about 15 miles west of Little Rock. Precambrian rocks are generally thought to be at considerable depths beneath the allochthonous Ouachita Mountains of Arkansas.

The foregoing data leads the authors to conclude that the early veins containing dickite at the Kellogg mines are probably very Late Pennsylvanian and Early Permian in age, and that the later pyrophyllite-cookeite-bearing veins are likely caused by ensuing very low-grade regional metamorphism in Early and Middle Permian time or slightly later. Finally, it is suggested that the primary mineralizing solutions at the Kellogg mine were derived from deep sources of unknown nature. Migration of these fluids was controlled by major fault and shear zones, and as such the enclosing sedimentary rocks yielded elements less materially to these veins than to the younger pyrophyllite-cookeite-bearing veins.

## ECONOMIC CONSIDERATIONS

Due to poor exposures and general lack of reliable geologic reports only a few observations can be made concerning the economic potential of this district. The silver content of the minerals is the primary reason that the Kellogg mines are of economic interest. For a mining operation to succeed all ore minerals (including both sulfides and oxidized materials) probably would have to be mined, concentrated and processed. Enrichment with depth of some of these metals is possible but unproven.

Initial exploration should be based on a sound geochemical, geophysical, drilling, and trenching program in an attempt to determine the tenor and extent of the mineralization. A reconnaissance ground magnetometer survey was made in the area by Commission geologists, but results were inconclusive.

The mineralized belt may be wider than now recognized and ore may occur beneath the extensive alluvium along Kellogg Creek to the north and also in the subsurface south of the known deposits. Some ore bodies may occur southward along the postulated cross faults (Fig. 4) and indeed, some of the presently known occurrences may be controlled, in part, by them. Subsurface prospecting immediately south of the known deposits seems logical because most of the observed veins dip south. Some exploration appears to be warranted a short distance to

the east and west of the district because of a similar geological setting.

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View of the Kellogg mill and ore shed in the late 1930's. The main shaft was located near the small building on the right.

# GEOCHEMISTRY, GEOTHERMOMETRY AND MINERALOGY OF COPPER, LEAD, ZINC, AND ANTIMONY DEPOSITS OF SEVIER COUNTY, ARKANSAS

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## ABSTRACT

Mineralization in the antimony district of Arkansas is characterized by Cu-Pb-Zn and/or Sb bearing quartz veins in east trending shear zones within the Mississippian age Stanley Group. Four stages of mineralization are recognized and include (1) an initial quartz-chlorite-chalcopryrite stage, (2) a younger sphalerite-galena-chalcopryrite stage, (3) a still younger arsenopyrite-jamesonite-zinkenite stage, and finally, (4) a stibnite-ankerite-calcite stage.

Homogenization temperatures from quartz and sphalerite show that mineralization occurred at minimum temperatures ranging from 115°C to 165°C for the first two stages, from 315°C to 206°C for the third stage, and from 214°C to 148°C for the fourth and final stage. These hydrothermal fluids were probably under pressures equivalent to a depth of burial of 3 km to 6 km.

Mineralogical, geochemical, and geothermometric data indicate different origins for the Cu-Pb-Zn mineralization and the younger Sb deposits. The former reflects rather low temperature hydrothermal activity related to unknown deep-seated sources during Pennsylvanian time, whereas the latter resulted largely from mobilization of formation waters of lower Paleozoic sediment during Late Pennsylvanian or Permian metamorphism.

## INTRODUCTION

Selected Cu-Pb-Zn deposits and Sb deposits of Sevier County, Arkansas have been examined to determine their origin, age, and structural relationship to the Ouachita Mountains of western Arkansas. The study involved brief visits to many mines and prospects to collect rock and mineral samples for laboratory study, and to observe the relationship of mineralization to structures preserved in the country rock.

The Arkansas antimony district, as the area has been named by previous workers, and its associated Cu-Pb-Zn deposits, is situated on the southern flank of the Ouachita Mountains (Athens Plateau portion) about nine miles north of DeQueen, Arkansas near the town of Gillham. Mineralization in the district occupies a belt about nine miles long and two miles wide that trends N.70°E. More than 20 mines and prospects were visited and the name and location of each is shown on Figure 1.

## PREVIOUS WORK

Early reports dealing with the Arkansas antimony district were by Williams (1875), Santos (1877), Dunnington (1878), Waite (1880), Ashley (1887), Comstock (1888), Jenney (1894), Bain (1901), Hess (1908), Schriver (1917), and Mitchell (1922). With the exception of Comstock's work in 1888, most earlier workers made only brief mention of the geology in the area. In 1923 Honess defined what appears to be a westward extension of this mineralization into McCurtain County, Oklahoma

when he described the geology and paragenesis of several Cu-Pb-Zn mines in that area. Both Miser and Purdue (1929) and Hall (1940) added considerable information about the general geology and mineralogy of the Arkansas antimony district. The first detailed report covering the history of mining activity, production figures and evaluation of the future economic potential of the area was given by Thoenen in 1944. McElwaine, in an unpublished report in 1952, provided additional geologic data on the Otto mine. An epithermal metallogenic province of Upper Cretaceous age was delineated by Scull in 1959 in which he included the antimony district. The two most recent works are by Stroud and others (1969) and Pittenger (1974).

## HISTORY OF PRODUCTION

The first mining operation was in 1842 with limited production of Cu, Pb, and Zn from the Davis mine. Next the Bellah mine was opened and produced lead for the Confederacy during the Civil War. It was not until shortly after the Civil War that antimony was accidentally discovered by a group of hunters who used some ore for the construction of a hearth. Following this discovery, the Wolfston mine was opened in 1874 and the Antimony Bluff mine in 1876. A small zinc smelter was constructed near Gillham during this time period and led to the development of the May, Otto, Valley, and Stewart mines.

Mining activity continued sporadically until 1947. Production was controlled by fluctuating prices generated by supply and demand of Cu, Pb, Zn and Sb. However all local mining activities were hampered by lack of transportation, dewatering equipment, and smelting facilities. Production records to date, as supplied by Stroud and others (1969), include 5388 tons of antimony ore and unknown amounts of copper, lead and zinc. Because nearly all mines are presently flooded, caved or otherwise inaccessible it is not possible to estimate the extent

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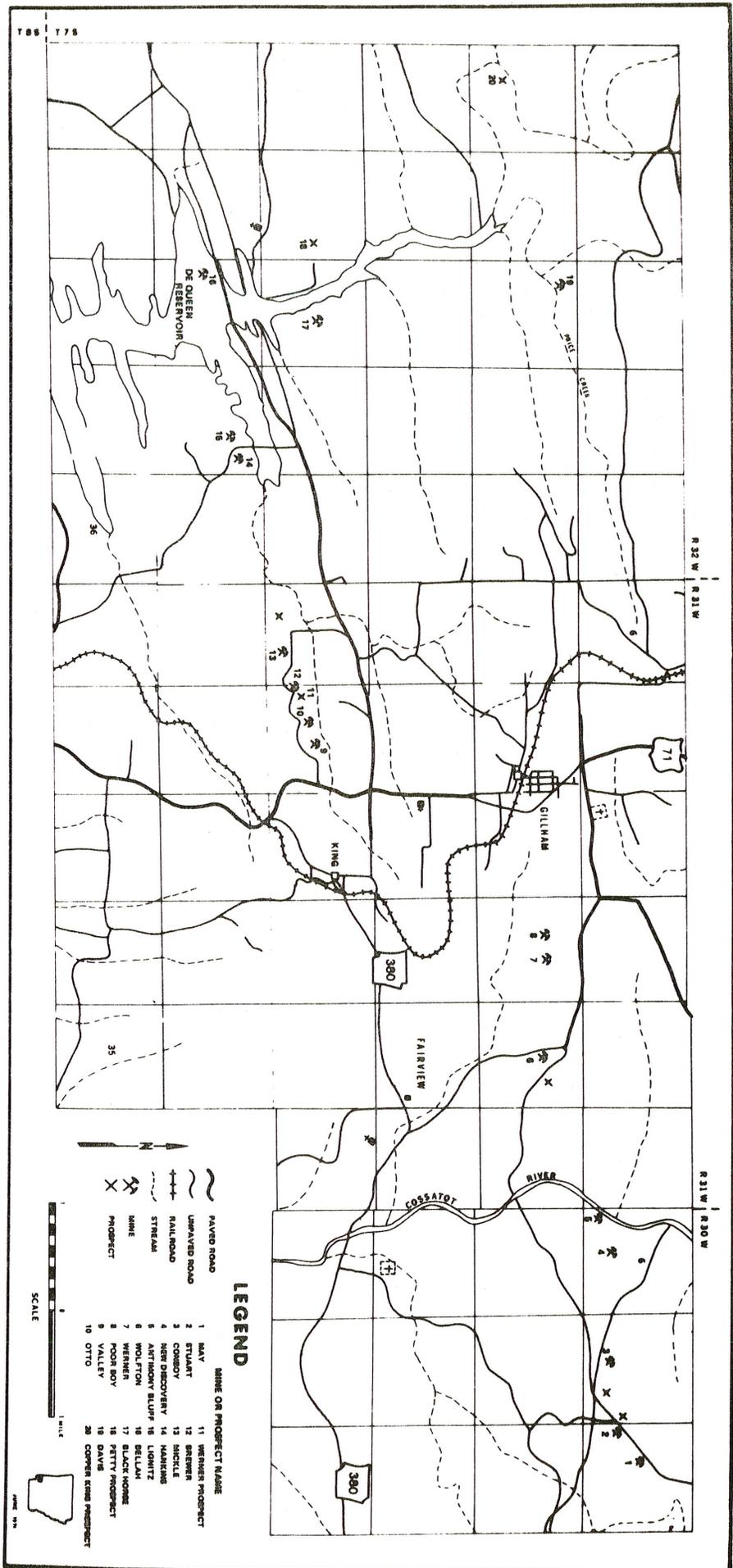


Figure 1. . . Mine location map of the antimony district of Sevier County, Arkansas.

of mining, except to say that it was not extensive either laterally or vertically.

## GEOLOGIC SETTING

Surface rocks surrounding and including the antimony district were originally assigned to the Standley Shale Formation by Taft in 1902. Modern usage has changed the name Standley to Stanley and resulted in elevating the Stanley Formation to the Stanley Group. The Stanley Group now consists of three Formations, named from oldest to youngest, the Ten Mile Creek, the Moyers, and the Chickasaw Creek (Harlton, 1938; Waithall, 1967).

Rocks in the Stanley range from about 6000 feet thick in western Arkansas to nearly 12,000 feet farther west in Oklahoma (Goldstein, 1959). According to Cline and Shelburne (1959) the Stanley and younger deformed rocks in the area are similar to the flysch shale facies of the Alps and Carpathians of Europe. This correlation was made because of the existence of convolute bedding, graded contacts between adjacent sandstone and shale, abundant flow and groove casts, and the scarcity of cross-bedding, ripple marks, and fossils except for planktonic and nektonic forms. Hickox (1968) established that sediment in the Stanley was the product of deposition from decelerating turbidity currents, particularly where graded bedding is observable.

In the vicinity of the antimony district the Stanley consists of gray-green to black laminated shale, which weathers buff, and medium to poorly sorted, angular to subrounded arkosic sandstone to platy siltstone. The feldspar content and average grain size of the sandstone increases towards the west. Individual beds vary from a few feet to as much as 100 feet in thickness.

The structural setting of the area was described by Honess in 1923 (McCurtain County area of Oklahoma) and Miser and Purdue in 1929 (DeQueen and Caddo Gap area of Arkansas). Both reports refer to a complexly folded and faulted anticlinorium (Benton-Broken Bow uplift) lying generally to the north of the zone of Cu-Pb-Zn-Sb mineralization. The core area of the anticlinorium exposes somewhat metamorphosed pre-Stanley rock having a total thickness of at least 3500 feet (Honess, 1923). Abundant east trending drag folds, reverse faults, crush zones, and quartz veins are characteristic of the area. Quartz-orthoclase pegmatites and a diorite sill have been reported by Honess (1923) nearby in Oklahoma. Near Gillham, Arkansas, Mitchell (1922) describes a five foot thick igneous dike that is probably genetically related to the many post tectonic lamprophyre dikes located farther east in Arkansas.

Cu-Pb-Zn-Sb mineralization is largely structurally controlled. This is evidenced by Sb veins at the Antimony Bluff mine that follow the N.75°E. trend of a large high angle reverse fault and also occur in several extension fractures generated by secondary strike-slip movement

along this fault. At the New Discovery mine Cu-Pb-Zn minerals intergrown with quartz and carbonate minerals follow a similar trend. The presence of gouge zones and secondarily developed extension fractures also indicate fault control of mineralizers. Other faults associated with ore are described by Thoenen (1944) at the Brewer mine and by McElwaine (1952) at the Otto mine. Even Comstock (1888) and Schriver (1917) recognized the importance of faulting when they suggested that mineralizers might have been guided along bedding as a result of bedding plane faulting. Conclusions drawn by the above workers are supported by the linear distribution of mines and prospects in the area. Examination of Figure 1 clearly shows this linear trend.

## MINERALIZATION

Mineral deposits in the Arkansas antimony district consist of four stages of mineralization, each separated from the other by a period of fracturing, and each distinguishable from the other on the basis of mineralogy, geothermometry, and age. Mineral identification was made by use of standard megascopic, microscopic, and x-ray diffraction techniques, and geothermometric data was obtained with the aid of a calibrated Leitz 350 heating stage on liquid filled vacuoles observed in polished thick sections of clear quartz, milky quartz, and sphalerite. The four stages of mineralization are discussed below.

### Quartz-Chlorite-Chalcopyrite Stage

The first and oldest stage of mineralization occurs along east trending fractures associated with the Ouachita orogeny. Quartz, chalcopyrite and accessory chlorite mark this stage of mineralization despite the fact that these same minerals may occur in somewhat younger Cu-Pb-Zn veins in the district. Quartz is commonly milky and generally nucleated around fragments of adjacent country rock. It was deposited throughout the entire stage of mineralization and deposition may have even continued uninterrupted into the second stage of Cu-Pb-Zn mineralization described below. Chalcopyrite is less abundant than quartz and often occurs near the base of quartz crystals. Both chalcopyrite and chlorite ceased to form following a second period of fracturing and structural adjustment that preceded a second more pervasive stage of mineralization. The paragenetic sequence for minerals associated with this stage of mineralization is shown in Figure 2.

Only one quartz crystal from the Bellah mine contained partially filled vacuoles that were usable in determining homogenization temperatures, and hence crystallization temperatures for this stage of mineralization. An average homogenization temperature of 152°C (uncorrected for pressure) was obtained for this material. Unfortunately the quartz crystal was taken from a small vug in a vein, and as such represents only that temperature which prevailed toward the end of this stage of mineralization.

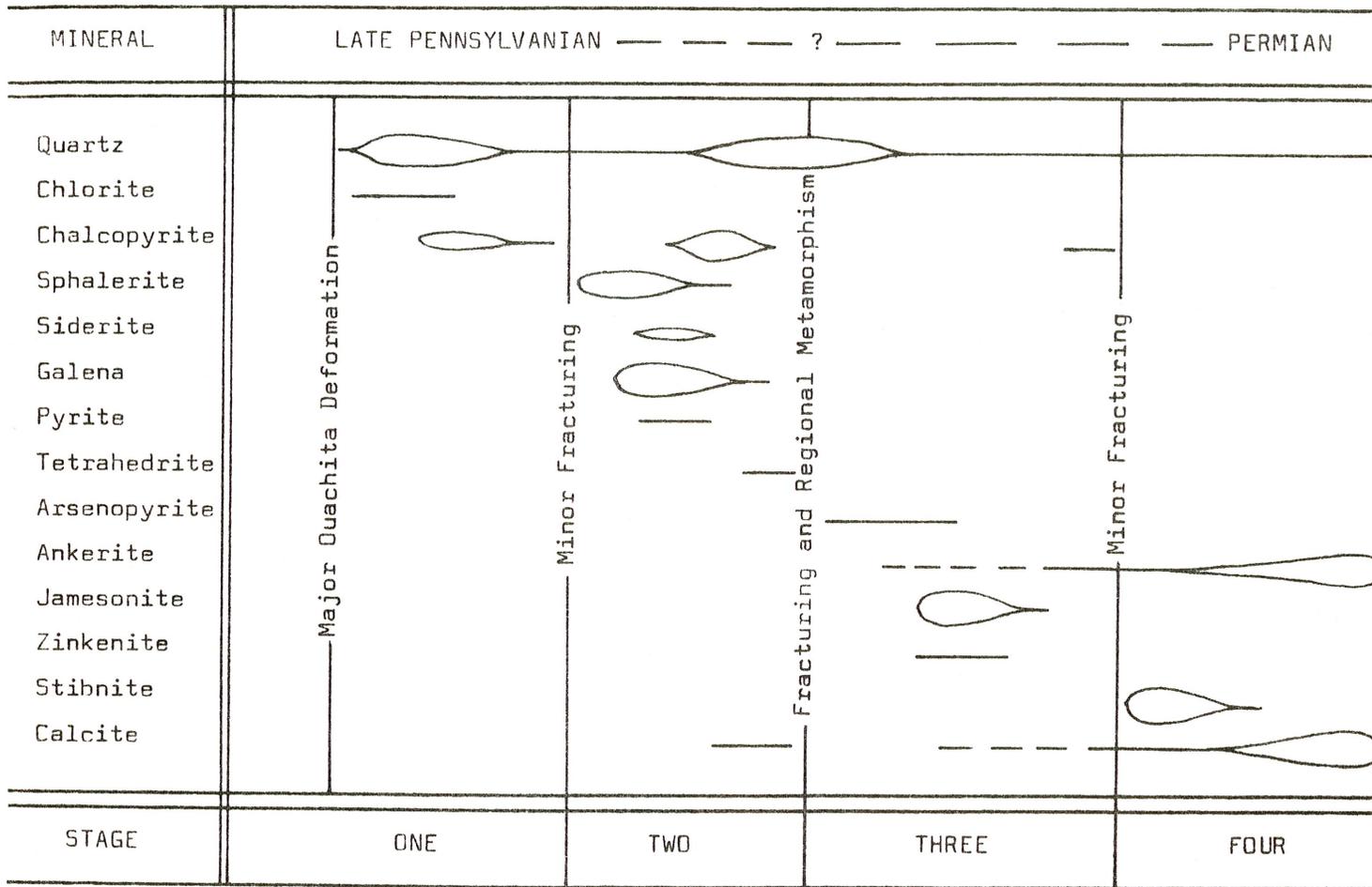


Figure 2. -- Composite paragenetic sequence of mineralization indicating stage of ore deposition and periods of structural and metamorphic activity.

### Sphalerite-Galena-Chalcopyrite Stage

The second stage of mineralization began following minor structural adjustments along older fractures that had been involved in stage one mineralization. Consequently older veins with previously formed quartz, chalcopyrite, and chlorite were reopened and younger mineralizers deposited sphalerite, galena, chalcopyrite, pyrite, siderite, and accessory calcite and tetrahedrite. Careful study of polished ore specimens from the Bellah, Davis, and New Discovery mines shows that these younger minerals consistently show crosscutting relationships with the older quartz, chalcopyrite, and chlorite of stage one. The order of deposition for these minerals is depicted in Figure 2 and chemical analyses of several samples from the Bellah mine are given in Table 1.

According to Bain (1901), a well defined three to eight foot thick vein at the Bellah mine cuts black shale of Carboniferous age. Galena is the dominant ore mineral above the 50 foot level and sphalerite with some galena and chalcopyrite is characteristic of the 115 foot level. Abundant slickensides with both vertical and horizontal striations, angular fragments of country rock surrounded by quartz, and comb and ribbon structure are noted. All subsurface data indicates the primary vein

trends nearly east, extends to a depth of at least 160 feet, and is associated with a major fault or shear zone. Unfortunately none of the subsurface workings are accessible and the surface area will soon be covered by waters of the DeQueen Reservoir. The Davis mine, as described by Allen (1910), Hall (1940), and Stroud (1969) was developed on a steeply inclined, east trending quartz rich vein up to 30 feet thick. A total of four shafts developed to a maximum depth of 204 feet, and several drifts up to 400 feet in length produced ore with 15 to 31 ounces of silver per ton and a trace of gold. The present study verified the presence of these metals along with minor antimony that is undoubtedly associated with accessory tetrahedrite. At the New Discovery mine, where a 100 foot shaft was sunk in 1902, similar mineralogical and structural patterns are present, except that ankerite is the most abundant gangue mineral. Other Cu-Pb-Zn occurrences are described by Honess (1923) in McCurtain County, Oklahoma that are probably genetically related to the Cu-Pb-Zn occurrences discussed above.

Samples of quartz and sphalerite taken from the Bellah and Davis mines were examined for fluid inclusions. The average homogenization temperature for primary fluid inclusions in quartz is 135°C (range 117°C to 142°C) from the Bellah mine and 139°C (range 133°C to 141°C)

TABLE 1. -- CHEMICAL ANALYSIS OF SELECTED MATERIALS FROM THE BELLAH MINE.

Data supplied by the Arkansas Geological Commission and results expressed in percent.

Element	Grab Sample	Galena	Chalco-pyrite	Siderite	Sphalerite
Ag	—	0.021	0.003	0.005	0.007
Cd	—	0.001	nil	0.001	0.130
Cu	6.80	0.700	28.200	0.158	0.068
Ge	—	—	nil	nil	0.002
Hg	0.05	0.070	1.120	0.070	0.100
Pb	0.38	44.000	0.089	0.030	1.080
Sb	0.03	0.121	0.054	0.044	0.051
Zn	36.30	4.290	0.017	0.105	57.400

from the Davis mine. Sphalerite from the same mines gave average temperatures of 133°C (range 115°C to 148°C) and 126°C (range 122°C to 128°C), respectively. At the New Discovery mine, however, sphalerite yielded a single filling temperature of 165°C. Homogenization temperatures from quartz from this mine were consistently above 250°C and are related to a younger stage of mineralization. The effects of this younger, higher temperature mineralization on stage two minerals is evidenced by partial recrystallization, deformation, and the presence of numerous vacuoles in sphalerite and quartz that appear to have been ruptured, allowing the liquid phase to escape. This situation is prevalent at all mines containing Cu-Pb-Zn that have been effected by younger, higher temperature mineralizing solutions.

#### Arsenopyrite-Jamesonite-Zinkenite Stage

A third stage of mineralization occurred following a period of fracturing and the on-set of subjacent Late Pennsylvanian to Permian<sup>3</sup> regional metamorphism associated with renewed arching of the Benton-Broken Bow uplift. Hydrothermal fluids heated to about 315°C moved upwards along fractures, some of which had served

as channelways for previous mineralizers that formed stage one and stage two mineral deposits. This mineralization began with the introduction of barren quartz and was followed by more quartz associated with arsenopyrite, jamesonite, zinkenite, pyrite, calcite, ankerite, and minor chalcopyrite. The paragenetic sequence for minerals associated with stage three mineralization is given in Figure 2.

Many mines in the antimony district contain minerals formed during this stage of mineralization. Although the Valley, May, Brewer, Wolfton, and Otto mines were producers of stibnite, all contain minerals deposited during stage three mineralization. This is possible because stibnite, a stage four mineral, was introduced along the same channelways as stage three minerals. The Valley mine operated in 1887-1888 and in 1916, and according to Hall (1940) consisted of a series of shafts and pits along a four to five foot wide vein that strikes N. 85° E. and dips 75° S. The presence of gouge indicates that the vein is associated with faulting. Stroud (1969) states that the May mine operated during 1887-1888, 1902-1903, 1916, and 1942 and produced 1510 tons of antimony concentrate. Three shafts, one as deep as 125 feet, driven in black shale and abutting against a massive sandstone encountered mostly younger (stage four) stibnite in veins up to 10 inches thick (Hall, 1940). Again the N. 80° E. trend, microbrecciation, and gouge indicates that ore fluids were guided by faulting. Both the Brewer and Wolfton mines have a geologic setting similar to mines described above.

The Otto mine, which operated in 1877-1888, 1916,

<sup>3</sup> Age of metamorphism as determined by Denison and others (see paper in Volume 1) using the K/Ar method on mica from metamorphosed samples from the test well Viersen and Cochran 25-1 located in sec. 25, T. 5 S., R. 23 E. of McCurtain County, Oklahoma. Total depth of well is 10,019 feet.

1937, and 1941 and was dewatered for geologic evaluation in 1952, is one of the best known mines in the area. Although 2500 tons of antimony concentrate was produced, sphalerite, galena, arsenopyrite, jamesonite, zinkenite, stibnite, ankerite, and calcite minerals are all abundant in the mine dumps. Therefore stages two, three and four are represented at this mine. According to McElwaine (1952) dewatering of the Otto mine revealed a 173 foot deep shaft that penetrated about 140 feet of black shale and then 33 feet of massive sandstone. Although little ore remained in place, stoping and drifting showed that most of the antimony production came from above the 115 foot level and that sphalerite, pyrite, quartz, and possibly some minor stibnite occurred below this level. In all cases production seemed to be associated with a mineralized quartz vein developed at the fault contact between black shale and massive sandstone.

The temperature at which minerals associated with stage three mineralization formed could only be determined by use of homogenization temperatures obtained from fluid inclusions in quartz that is genetically related to arsenopyrite, jamesonite, and zinkenite. Quartz associated with jamesonite at the Wolfton mine gave homogenization temperatures of 242°C to 252°C (average 247°C), at the Valley mine 249°C to 253°C (average 251°C), and at the Brewer mine 206°C to 227°C (average 217°C). However some barren quartz believed to be related to the earliest phase of this third stage of mineralization gave filling temperatures as high as 315°C. Thus it may be concluded that early barren quartz began forming at 315°C and as the temperature declined to about 250°C jamesonite and other ore minerals began to form until temperatures declined to around 215°C at which point most stage three mineralization ceased.

#### **Stibnite-Ankerite-Calcite Stage**

The final stage of mineralization began following a Late Pennsylvanian or more likely a Permian period of fracturing. Stibnite, ankerite, calcite, and some quartz characterize this stage of mineralization and show the crystallization sequence given in Figure 2. The best example of stage four mineralization occurs at the Antimony Bluff mine where stibnite fills N. 20° E. trending gash fractures developed in sandstone that are related to strike-slip movement along an older N. 75° E. trending reverse fault. Mining activity in 1876 at this mine produced about 10 tons of antimony concentrate. Waite (1880) reports that ore bottomed-out at a depth of 55 feet.

Stibnite was the only ore mineral to develop during stage four mineralization. Thus stibnite occurring with other ore minerals indicates more than one stage of mineralization. The coexistence of ore minerals and their associated gangue minerals not only complicates the paragenetic sequence of mineralization, but make geothermometric determinations more hazardous. Nevertheless filling temperatures for vacuoles in vein quartz associated with stage four mineralization have been determined. Average

filling temperatures for quartz intergrown with stibnite at the Antimony Bluff and Valley mines is 214°C. At the May mine quartz intergrown with both stibnite and ankerite gave an average temperature of 188°C. And finally quartz younger than stibnite and equal in age to very young ankerite yielded a filling temperature of 148°C.

Analysis of the above temperature data indicates that stage four mineralization began at temperatures as high as 214°C and ended when the temperature of the ore fluids declined to 148°C. Stibnite crystallized over a temperature range of 214°C to 188°C. As in all previous sections, temperatures are uncorrected for pressure.

### **GENESIS OF ORE FLUIDS**

Copper, lead, and zinc deposits, which developed over a temperature range of 115°C to 165°C, have been shown to be the oldest mineral deposits in the area. Hydrothermal fluids genetically related to these deposits were guided by shear zones and related fractures formed during early (Pennsylvanian) stages of deformation associated with the Ouachita orogeny. Although igneous rocks are uncommon to the area, the authors suggest that some deep-seated igneous source must have supplied most of the Cu, Pb, and Zn in these deposits. This conclusion is supported by chemical studies on Paleozoic shale and sandstone that could be considered as potential source beds for the metals (Table 2). It is clear from this limited study that neither the sandstone nor the shale tested contain anomalous Cu, Pb, or Zn.

Antimony deposits characterized by stibnite, jamesonite, zinkenite, and arsenopyrite formed over a temperature range of 315°C to 148°C, and are among the youngest of the metal bearing veins in this area. Hydrothermal fluids that generated these deposits certainly obtained some metals (Cu, Pb, Zn) from older deposits by resolution of previously deposited ore minerals. The authors suggest, however, that formation waters associated with subjacent sediments probably contributed significantly to the supply of both Sb and As in these deposits. Table 2, which shows anomalous Sb and As in nearby shale and sandstone supports this thesis.

#### **Barren Quartz Veins**

Many barren quartz veins occur in the antimony district as well as throughout the entire Ouachita Mountain region and have been the object of much study (Miser, 1943 and 1959; Engel, 1946 and 1952). The relationship between barren quartz veins and quartz veins containing Cu, Pb, and Zn has never been fully determined. The primary physical differences between these two types of veins are; (1) the barren veins are generally thicker, (2) they do not contain numerous wallrock fragments, (3) they are not as extensively fractured and slickensided, and (4) they generally reflect the composition of the enclosing wallrock. These observations indicate that the bulk of barren quartz formed more recently than Pennsylvanian quartz veins

TABLE 2. -- COMPARISON OF AVERAGE SHALE AND SANDSTONE (TUREKIAN AND WEDEPOHL, 1961) WITH THAT OF UNMINERALIZED STANLEY SHALE (GC-7) AND SANDSTONE (GC-9) IN ANTIMONY DISTRICT. (Analysis by U. S. G. S. and results expressed in parts per million.)

Element	Shale GC-7	Sandstone GC-9	Average Shale	Average Sandstone
Fe	70,000	7,000	47,200	9,800
Mg	10,000	2,000	15,000	7,000
Ca	1,500	1,000	22,100	39,000
Ti	5,000	2,000	4,600	1,500
Mn	500	200	850	150
As	N (100)	500	13	1
B	150	50	100	35
Ba	1,000	150	580	250
Be	7	1	3	---
Cd	N(2)	N(2)	0.3	---
Co	20	5	19	0.3
Cr	150	20	90	35
Cu	30	10	45	10
La	70	30	92	30
Mo	N(5)	N(5)	2.6	0.2
Nb	N(10)	N(10)	11	---
Ni	50	5	68	2
Pb	20	5	20	7
Sb	50	100	1.5	0.4
Sc	20	N(5)	13	1
Sn	7	N(5)	6	---
Sr	100	N(100)	300	20
V	200	20	130	20
W	N(50)	N(50)	1.8	1.6
Y	30	15	26	40
Zn	100	N(50)	95	16
Zr	200	500	160	220

N ( ) indicates element not detected at limit of detection indicated within parentheses.

containing Cu, Pb, and Zn minerals, and according to Bass and Ferrara (1969) are Late Pennsylvanian to Early Permian in age. However, if the younger antimony bearing quartz veins are related more closely to the thermal effects of Permian metamorphism, then these mineralized quartz veins would be among the youngest reported in western Arkansas to date, and may therefore be younger than some barren quartz veins.

## Oxidation Products

Most of the ore minerals that lie within the zone of oxidation have been partially altered to form secondary minerals. Secondary minerals include limonite, hematite, azurite, malachite, covellite, cerussite, anglesite, smithsonite, bindheimite, cervantite, and stibiconite. In general oxidation does not seem to be extensive and probably did not extend more than 30 feet below the present surface of erosion.

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# GEOCHEMICAL SURVEY OF THE SILVER, ARKANSAS AREA<sup>1</sup>

By John J. Santos<sup>2</sup>

## ABSTRACT

Soil samples from the area of Silver, Montgomery County, Arkansas, were analyzed for relative heavy metal (lead, zinc, and copper) content. Several geochemical anomalies were found. All of these anomalies are associated with abandoned mines or prospect pits with the exception of a small anomaly west of the town of Silver and part of an anomaly in the Tompkins Bend area. Trend-surface analysis did not show up any additional anomalies.

## INTRODUCTION

The purpose of this study was to collect soil samples, analyze the samples for relative heavy metal content (lead, zinc, and copper) and interpret the results of the analyses with the aid of the IBM system 370 computer. The samples were collected in an area of approximately eight square miles in the vicinity of the community of Silver in Montgomery County, Arkansas (Figure 1). This area was chosen because it is known to contain lead and zinc sulfide deposits that could be compared with anomalies found by this study.

Little published material is available on the geology of this area. T. B. Comstock (1888) included the Silver area in his work on the mineral resources of west-central Arkansas. The area is also referred to briefly by Anderson (1942), Engel (1951), and Stroud and others (1969) who reported the occurrence of metal sulfides.

Geochemical prospecting is based on the principle that the distribution of some element of an ore deposit will form a distinctive pattern in the unweathered country rock containing the ore body or in the rock's weathering products. Often this distribution pattern is formed by valuable metals from the ore minerals. However, associated metals, called trace elements, can also form important distribution patterns. These distributions are referred to as geochemical halos or anomalies and are divided into two groups: primary and secondary. Primary anomalies are a bedrock feature, not related to erosion. They are the result of dispersion into the country rock around the ore body by the mineralizing fluids that formed the ore body. Anomalies formed by the concentration of ore minerals during consolidation or lithification would also be considered primary anomalies. Secondary anomalies are erosional-depositional features. They form from dispersion resulting from weathering of either the ore body or its primary anomaly.

## GEOLOGY

The country rock underlying the area of study is entirely Womble Shale, named for the town of Womble (now Norman), Arkansas. The Formation consists primarily of shale with a few thin beds of sandstone near the base and limestone near the top (Pitt, et al., 1961, p. 80). Rarely did the author see beds of sandstone in the Silver area. No limestone was seen.

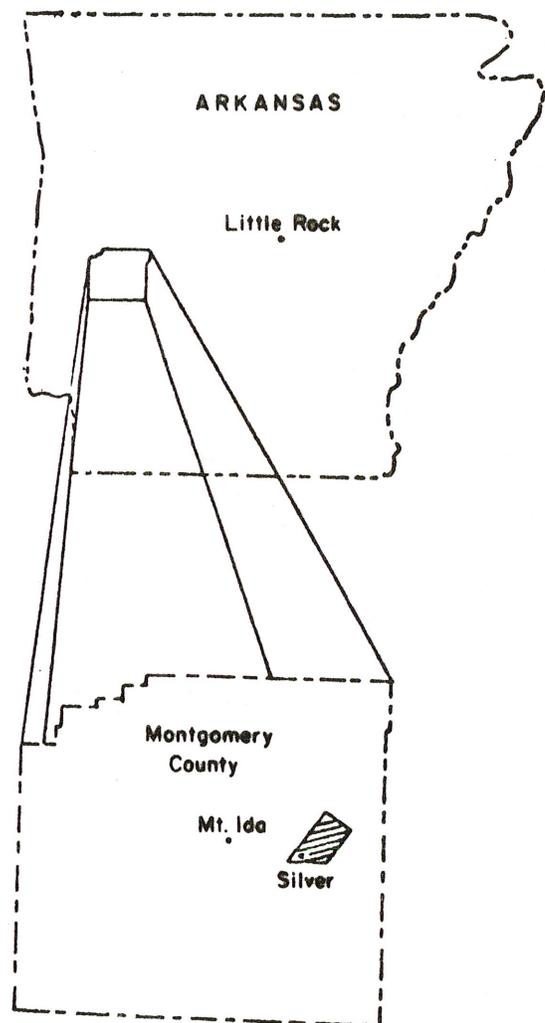


Figure 1. -- Location map.

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The Womble is conformable with the Blakely Sandstone above and the Bigfork Chert below. Two graptolite faunas have been found in the Womble, one early Middle Ordovician in age and the other late Early Ordovician (Croneis, 1930, p. 99).

The shale in the Silver area is black, graphitic, hard and fissile. It weathers to tan or reddish brown and breaks into angular blocks often rhombic in shape. The sandstone is dark gray, fine-grained, hard and well cemented, and weathers to a tan color.

The Silver area is less than a mile north of the west-southwest to east-northeast trending axis of the Ouachita Anticlinorium (Croneis, 1930, Plate 1-B). In this position it would be expected that the rocks in the area would strike parallel to the axis of the anticline and dip to the north. Because of the soil cover and thick vegetation in the area, there are few outcrops where strike and dip could be measured; and on the outcrops where they could be measured they were inconsistent and changed considerably within a few feet.

Since the Womble Shale is incompetent, the Ouachita orogeny produced numerous small folds in the Formation. In the Silver area several tightly folded zones were seen. The direction of plunge of the fold axes could be measured at only a few locations, and the pattern formed by these fold axes is inconclusive. In most of the area the folding is gentle with wave length greater than amplitude. This results in gently dipping beds.

Veins of quartz are common in the area and are thought to be controlled by structure (Miser, 1943, p. 98).

## METHODS OF INVESTIGATION

### Field Methods

A grid was drawn on a plastic overlay to fit over an aerial photograph (with a scale of 2.85 inches per mile) of the area so that a sample could be collected approximately every quarter mile. In those parts of the area to which access was limited fewer samples were collected, and additional samples were collected in areas that showed unusually high metal content. Where the soil profile was exposed in road cuts or by erosion along the shore of Lake Ouachita, one to one and one-half feet of material was removed to insure that leached soil was not sampled. Four to five grams of soil were collected from the B-horizon, two to three feet below the surface. At other locations a hole was drilled two to three feet below the surface with a hand auger and a sample taken from the soil adhering to the lower few inches of the auger.

A map of the area was made from aerial photographs. In addition to marking the location of the samples, the geology of the area was also mapped.

### Laboratory Methods

The samples were analyzed for heavy metal content using ammonium citrate. The method that was used is described by Bloom (1955) as a rapid method which will allow the heavy metal analysis to be made in the field. Comparison by Bloom of analyses using aqua regia and ammonium citrate for digestion of the samples shows ammonium citrate removes 1 to 30 percent of the available metals and he reported anomalies were clearly outlined in each case (Bloom, 1955, p. 35). A description of the methods of sample analysis used in this study can be found in Ward, et al., 1963, "Analytical Methods Used in Geochemical Exploration".

### Computer Analysis of Data

A trend-surface analysis was made of the heavy metal concentrations from 87 samples using a Fortran program developed by Good (1964) for the IBM 1620 and modified for use on the IBM 370.

With trend analysis local variations are separated from the regional variation by fitting a trend-surface to the observed field data and showing local variations from this surface. In this case a first degree (plane) surface through a fourth degree surface were computed.

## CONCLUSIONS

There are no production records of the mining activity in this area (Stroud and others, 1969). Mine locations that were found during this survey and those that are shown on maps of the area (Anderson, 1942, Plate 5 and U. S. Geological Survey, 15' Mt. Ida quadrangle, 1959) are shown on Figure 2. Since the area has been thoroughly prospected (Stroud and others, 1969) additional mines may have existed.

A relatively large anomaly shows up in the area of Tompkins Bend. The Montezuma mine is in the southwest part of this anomaly. A small prospect is located in the eastern part of the anomaly. Contamination by ore material brought up from the mine could account for part of this anomaly. However, since the mine is in the southern part of the anomaly, the anomaly terminus that is nearest to the only main road in the area, it is not likely that contamination could have caused an anomaly of this extent. The anomaly shows up as a residual high on all four degrees of the trend-surface analyses. It also shows up on the third and fourth degree analyses as a high on the trend-surface.

Smaller residual highs show up around the Minnesota and Waterloo mines and the unnamed mine in the western part of the area. The trend-surfaces do not show any significant highs in these areas.

A small residual high shows up in the southwest corner of the area along the main highway. There is also a high on

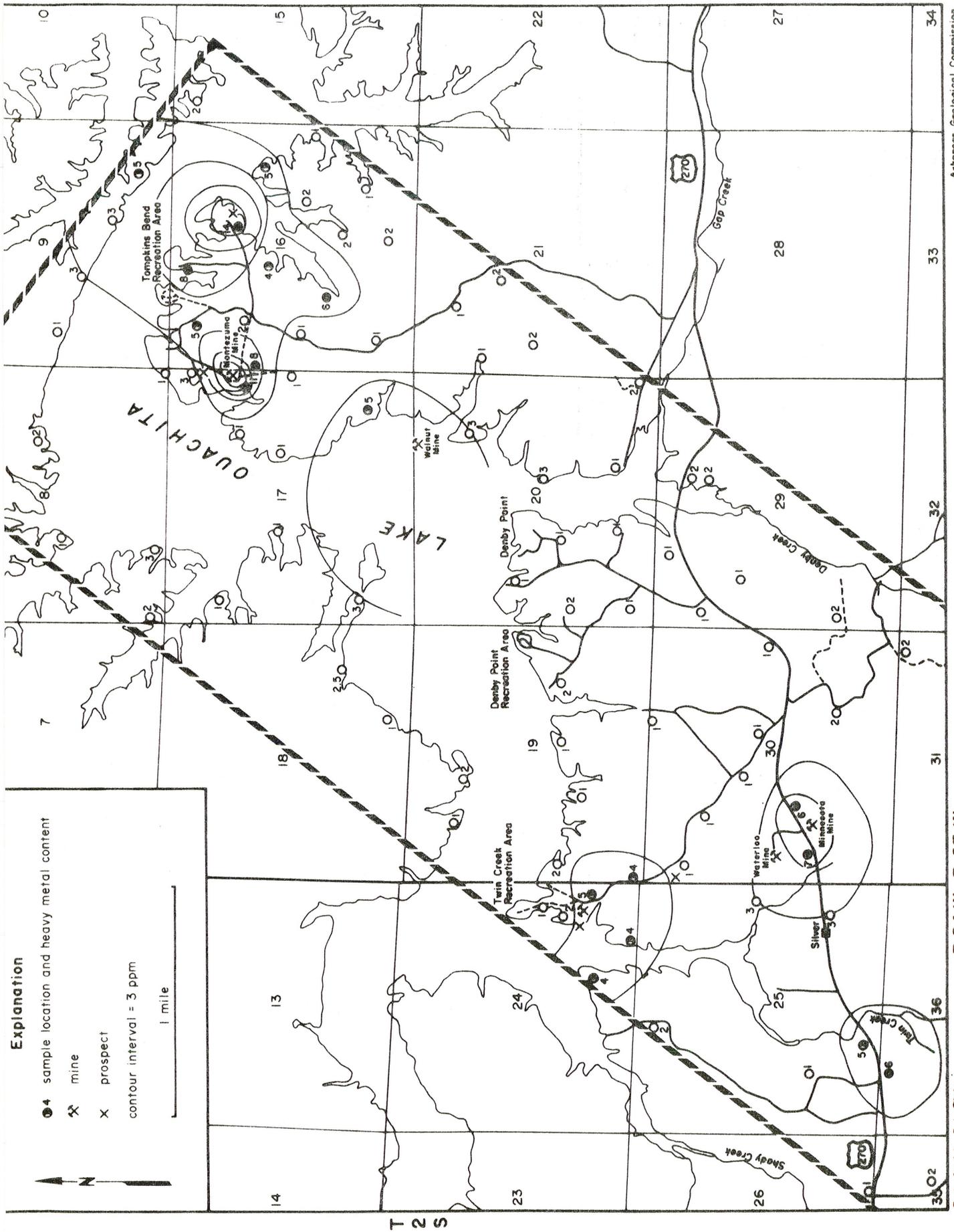


Figure 2. ... Geologic map of the Silver, Arkansas area showing the heavy metal content (lead, zinc, and copper).

the third degree trend-surface in this area. There is no reference to mining activity at this location in the literature and none was seen at the location.

The small residual high in the northwest part of the area on the second through fourth degree residuals is probably the result of fitting the trend-surface to the anomaly in the Tompkins Bend area. The heavy metal content of samples in this area is not above background level.

The anomaly in the Tompkins Bend area indicates the possibility of sulfide deposits northeast of the Montezuma mine. These deposits could be of economic interest if they are extensive enough. However, if they occur in small pockets similar to the deposits described by Comstock (1888), it is doubtful that they could be mined at a profit. The same condition is true for the small anomaly in the southwest part of the area. All other anomalies that show up surround mines that are known to be worked out (Comstock, 1888).

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**Editor's addendum**--thesis investigations during 1977-1978 by Andrew Kurrus of the University of Arkansas has shown that some of the abandoned mines and prospects in the Silver, Arkansas area were improperly located by some previous workers.

# ASPECTS OF MANGANESE PRODUCTION IN WEST-CENTRAL ARKANSAS

By Raymond B. Stroud<sup>1</sup>

## ABSTRACT

This report by the Bureau of Mines describes briefly the manganese district of west-central Arkansas and presents information on certain manganese deposits that were the sites of manganese production during the most productive period 1955-59.

## INTRODUCTION

Under current technology employed by steel manufacturing processes in the United States, manganese is an essential component in making cast iron and steel (11).<sup>2</sup> More than 90 percent of the manganese used annually in the Nation is consumed for metallurgical purposes. The United States has long imported the major part of its manganese needs from foreign sources, and in 1971 this country imported nearly all of the manganese consumed by industry (5).

Because of the essential use of manganese, the apparent lack of economic reserves in the United States, and the dependency on foreign imports, the Federal Bureau of Mines has from time to time investigated, in varying intensity, manganese resources in this country. One such study was made in 1959 in west-central Arkansas where, at the time, manganese production had reached its highest level since discovery in the mid-19th century. This work is the principal basis of the present report.

The Bureau of Mines conducted a reconnaissance survey of manganese mining and processing activities in Polk, Pike, and Montgomery Counties, Arkansas, referred to herein as the West-Central District, in the spring of 1959, to determine the extent and nature of manganese production and to gain insight into potential manganese resources of the district (14).

The geology of the West-Central Arkansas Manganese District was first described in depth by R. A. F. Penrose (10) and later by H. D. Miser (8), and H. D. Miser and A. H. Purdue (9). Other subsequent reports (1-3) draw heavily on information provided by the aforementioned authors. The present report, in addition to details of the reconnaissance investigation, contains a summary of basic geological data.

The Bureau of Mines first investigated the district on a reconnaissance basis in 1955 (4) and results of a mineral-dressing study were published in 1956 (6). Manganese production potentials are incompletely known because of the lack of certain critical data, described later in the present report.

## ACKNOWLEDGMENTS

L. P. Kelone, cartographer, Arkansas Geology Division, Department of Commerce, Little Rock, Arkansas, is due special recognition for preparation of the illustration that accompanies this report.

Charles G. Stone and Drew F. Holbrook, geologists, Arkansas Geology Division, Department of Commerce, Little Rock, Arkansas, reviewed this report and made a number of helpful suggestions.

## GEOLOGICAL SUMMARY

### Location

The manganese district of west-central Arkansas as discussed in this report includes manganese deposits found in Polk, Montgomery, and northern Pike Counties, Arkansas. The mines that have been most productive are in an area south and east of Mena, Polk County, and in the southwest part of Montgomery County. The district trends east-west for about 35 miles and is from about 6 to 8 miles wide north to south (Fig. 1). Manganese occurs also in Garland, Hot Spring, Saline, and Pulaski Counties and these deposits have been described by Penrose (10). Production of manganese in these counties has been virtually nil.

### Geography

Manganese deposits of commercial significance occur mainly in the Arkansas Novaculite of Devonian and Mississippian age in Polk, Pike, and Montgomery Counties. A substantial tonnage of manganese ore was mined from a shale section in Montgomery County that originally was thought to be Stanley Shale of Mississippian age; however, later geologic work by Stone and Haley indicates the deposit is mostly an upper shale sequence of the Arkansas Novaculite (12). The formations have been tightly folded

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<sup>2</sup>Numbers in parenthesis refer to references at the end of this report.



into east-west trending belts and subsequent erosion has formed the rough, mountainous terrain of the Novaculite Uplift section of the Ouachita Mountains Province. Peaks and ridges rise from about 500 to nearly 1,500 feet above valley floors. Hillsides are steep and rugged and access is difficult north to south except through water gaps, saddles, or by steeply inclined switchback roads. Principal roads traverse the district east-west and travel is relatively easy.

Headwaters of the Cossatot, Saline, Little Missouri, and Caddo Rivers rise in the mountainous area and flow south and southeastward. These streams might provide adequate flows of water suitable for small milling operations in the area of interest.

Rail connections on the Kansas City Southern Railway are available at Mena, the county seat of Polk County, and largest town near the manganese district.

The rural areas apart from Mena and other much smaller settled communities are sparsely populated and much of the land comprises part of the Ouachita National Forest.

### Stratigraphy

Rocks containing manganese, including underlying formations that crop out in the area of interest, are of Paleozoic age and from oldest to youngest include Crystal Mountain Sandstone, Womble Shale, Bigfork Chert, and Polk Creek Shale of Ordovician age, Blaylock Sandstone, and Missouri Mountain Slate of Silurian age; Arkansas Novaculite of Devonian and Mississippian ages; and Stanley Shale of Mississippian age. Deposits of manganese that have been productive are chiefly in hard, dense Arkansas Novaculite and very infrequently in shale. The formations cropping out in the manganese district are described fully by Miser (8). Some essential characteristics of host formations are described in the following four paragraphs.

In the area of interest, the Arkansas Novaculite is comprised of three divisions. The lower and most of the middle divisions are of Devonian age and the upper division is of Mississippian age. The lower division consists mainly of highly siliceous massive, white novaculite, and the basal section contains interbedded shale. In the vicinity of some manganese occurrences and elsewhere, the beds are heavily shattered, and joints are sometimes filled with quartz. On occasion, the hard novaculite is softened at the surface apparently because of weathering. The division is about 150 to 300 feet thick in most localities, but in some places, it is as little as 25 feet thick and in others as much as 410 feet thick (8).

The middle division of the Arkansas Novaculite consists essentially of interbedded shale and novaculite, a conglomerate at the base of the division is present in many areas. The novaculite of the middle division appears similar to that in the lower division, but it is usually dark gray or black in color. Fresh shale in the division is black but weathers to usually lighter colors on the outcrop. A basal conglomerate consists mostly of novaculite pebbles cemented with sandy flint. The middle division changes

substantially in thickness ranging from about 75 to 575 feet from place to place. The middle division also exhibits many of the fractured characteristics developed in the lower part of the formation.

The upper division of the formation is typically massive novaculite, light gray in color, in part tripolitic and highly calcareous in places. It, too, is extremely variable in thickness, ranging from as little as 20 feet to as much as 125 feet. Because of its apparently high calcium carbonate content, which is lost on weathering, much of the upper novaculite division is soft and porous and weathers to lighter shades in some places. Close jointing is prevalent as in the other divisions. In some northern exposures, interbedded novaculite and siliceous shale predominates.

Characteristics of the Stanley Shale as described in this report deal only with the lowermost part of the formation and more particularly where manganese has been found. The Stanley is typically shaley and exhibits slaty cleavage. It is dark gray to black on most unweathered surfaces and light gray to buff on exposed surfaces; however, green, red, and brown shades prevail locally. Because of strong deformation, the shale beds are crumpled and contorted in the area under discussion. Lenticular bodies of sandstone nearly altered to quartzite and thin novaculite conglomerates are not unusual near the base of the formation. Quartz veins that appear to follow fracture systems are frequently found.

### Structure

Compressive forces from both the north and south have folded the rock strata in the West-Central Manganese District into relatively narrow belts that crop out along sinuous ridges and valleys that typify topography in the area. This folding has developed a substantial number of anticlines and synclines, and because of erosion, repetitive exposures of formations may exist north to south. In some instances, folding has overturned the beds to the degree that older strata crop out at the highest points of relief, and younger beds are topographic lows.

The folding has crumpled and contorted the shale beds intensely, and it is difficult to follow bedding planes over long distances. The more resistant rocks are heavily fractured, but bedding planes persist and can be traced with moderate difficulty.

Thrust faulting is commonplace, and jointing both parallel and oblique to bedding planes is ubiquitous. Joints tend to be healed in softer strata whereas fractures in hard rock stand open or are in some places filled. The fractured character of the hard beds has left passageways for ascending, descending, and lateral movements of water.

### Manganese Occurrences

Manganese ores mined in the West-Central Arkansas Manganese District include psilomelane, pyrolusite manganite, and wad. Minute quantities of braunite have been found in one known instance (6, p. 6). Of these,

the most important have been psilomelane and pyrolusite. Although information on complete chemical analyses is scant, several other elements are associated with the manganese including iron, phosphorous, aluminum, cobalt, calcium, barium, magnesium, potassium, sodium, copper, nickel, lithium, sulfur, and silicon. Other metalliferous minerals identified from the area of interest and in part associated with the manganese include limonite, native copper, cuprite, scant chrysocolla, turquoise, malachite, pyrite, and lithiophorite. Wavellite, variscite, dufrenite and other iron phosphate minerals, barite, and quartz occur selectively. Arsenic was reportedly found in one instance associated with a deposit of pyrite and marcasite (9, p. 175).

Two or more manganese minerals commonly occur together in the form of loose nodules and boulders, and in pockets and veins following fracture systems in novaculite. Manganese minerals as replacement and vein material are also found in a shale sequence at one mine (7).

Manganese-bearing veins are from less than 1 inch to as much as 4 feet thick, as observed by Penrose and Miser (8-10). Miser states that an 8-foot vein was reported at the North Mountain mine (8, p. 79). Heavily fractured zones in the novaculite that contain minable manganese are as much as 50 feet thick (13, p. 331). In these zones, the manganese acts as cement in the heavily brecciated novaculite.

Limonite and goethite, clay, and phosphate minerals accompany the manganese ores and at times occur separately. Frequently, the iron appears to cap the upper parts of a manganese ore body.

Little is known regarding the occurrence and exact association of other metallic minerals such as copper, cobalt, and nickel that have been identified in certain samples of manganese. Specific mineralogic and geochemical studies may be warranted because the copper content of some samples exceeds 1.0 percent and thus may be of economic interest.

Analyses of various samples of manganese minerals and ores are given in reports by Penrose, Miser, and Miser and Purdue (8-10). Table 1 contains analyses of representative samples of manganese ore studied in mineral dressing tests by Fine and Frommer (6).

### Origin

In attempting to evaluate potential commercial significance of manganese occurrences in west-central Arkansas, it would seem that origin is of primary importance. Miser and Purdue point out in one instance that manganese deposits associated with fractures zones are apparently separated vertically as much as 800 feet (9, p. 170). Because of a lack of exploration in this instance and others, where vertical separations are less, it is impossible to state that

TABLE 1. -- ANALYSES OF MANGANESE ORES FROM WEST-CENTRAL ARKANSAS DISTRICT

CONSTITUENT IN PERCENT	SAMPLE IDENTITY -- locations at end of Table							
	A	B	C	D	E	F	G	H
Manganese	15.9	19.7	4.4	5.6	24.0	24.0	26.7	14.1
Iron	10.1	1.7	2.8	.74	2.3	1.8	.96	.7
Silica	53.6	50.3	85.3	85.9	50.0	47.1	22.4	72.2
Calcium carbonate	4.0	---	---	---	---	---	---	---
Alumina	3.6	7.6	---	.39	2.7	1.8	18.4	.4
Phosphorus	.39	.11	.11	.10	.14	.19	.09	.05
Barium	---	---	---	---	---	.62	---	---
Copper	---	---	---	---	---	---	1.2	.06
Nickel	---	---	---	---	---	---	.64	.03
Cobalt	---	---	---	---	---	---	.20	.03
Lithium	---	---	---	---	---	---	.24	---

Sample A from property of Manganese Development Corp. in secs. 13 and 14, R. 28 W., T. 4 S., Polk County.

Sample B from North Mountain mine, SE $\frac{1}{4}$ NW $\frac{1}{4}$ , sec. 10, T. 4 S., R. 27 W., Montgomery County.

Sample C (sample IN-M-1) composite from several dozen trench samples from property controlled by Inland Mining Corp.

Sample D (sample IN-M-2) representative sample from 13-foot face of outcrop on the south side of American group of claims Inland Mining Corp.

Sample E from A. B. Pickell property in SW $\frac{1}{4}$ NW $\frac{1}{4}$ , sec. 11, T. 3 S., R. 30 W., Polk County.

Sample F (sample 1) from properties in T. 4 S., R. 30 W., controlled by Duncan, Dieckman and Duncan Mining Co.

Sample G (sample 2) from Duncan, Dieckman and Duncan Mining Co. property.

Sample H (sample 3) property of Duncan, Dieckman and Duncan.

(6).

the manganiferous zones are connected. But because of the apparent continuity of the fractures horizontally and vertically, it is possible that the mineralized zones persist. Penrose and Miser suggest that the origin of the novaculite and manganese are closely associated (8-10). They maintain that the manganese could have been deposited as finely disseminated material in the host novaculite and that this disseminated manganese was later concentrated by descending ground water passing through the vast amount of overlying rocks that have subsequently been eroded.

The author contends that evidence offered by Penrose and Miser does not rule out that the manganese deposits are most likely hydrothermal in origin. Mineralogical and geochemical studies have not been made in the area of interest on a systematic basis, and such research could offer further clues to origin. The mineral suite found in association with some manganese ores in the district that show the presence of copper, cobalt, nickel, lithium and possibly arsenic suggest the possibility of telethermal mineralization. In many instances, manganese minerals, particularly psilomelane, are intimately associated with quartz veins. When the manganese mineralization is considered regionally with nearby deposits of mercury, antimony, and copper-zinc-and lead-bearing quartz veins found in the Ouachita Mountain Province, a hydrothermal origin is conceivable. The filling of fractures and brecciated zones in novaculite with manganese, and, an apparent replacement and associated vein deposit of manganese in shale may have been accomplished through ascending solutions. Since original emplacement in the broken zones of novaculite, subsequent movement of the manganese, and redeposition with possible enrichment has probably occurred through ground water movements.

Until sufficient exploration is done and appropriate geochemical and mineral studies are made, the origin of the deposits must remain in question.

## HISTORY AND PRODUCTION

Following discovery of manganese, about the middle of the 19th century in west-central Arkansas, extensive prospecting in the latter half of the 1880's led to a small production. In 1886, 4 tons of manganese was produced in Polk County (North Mountain mine). This activity was followed by several periods of development and mining activity particularly under the stimulus of high prices during World Wars I and II and the decade of the 1950's. From the late 1880's until the end of 1955 about 2,000 long tons of manganese was reportedly produced. Most of the ore was mined in Polk and Pike Counties. Manganese purchases for the Federal Government National Stockpile program of the early 1950's led to significantly increased production in 1958 and 1959. This buying program ended in August 1959, and since that time no manganese production has been recorded.

In 1958 and 1959, six mines produced most of the ore;

three of these accounted for the major part of the production. Two concentrating mills were operated in 1958 and one mill was active in 1959. The 1958 reported production amounted to about 3,100 tons of concentrates and direct shipping-grade ore; the 1959 output was an estimated 4,100 tons of concentrates and direct shipping-grade ore.

Value of the more than 7,000 tons of ores and concentrates recorded as production in 1958-59 was about \$750,000.

## MINING AND PROCESSING

Trenches, pits, adits, and shafts have been dug at more than 50 separate sites in Montgomery, Pike, and Polk Counties. Scant exploratory drilling was done in the 1950's.

To illustrate the mining activities that prevailed in the 1950's, partial descriptions of significant mines follow. Tables 2, 3, and 4 present some available details concerning these operations and aspects of other mining activities in the area of interest.

### Mines

#### Lost Hatchet Mine

The Lost Hatchet mine was opened in 1954 in the SE¼ NW¼ sec. 36, T. 3 S., R. 28 W., Polk County (Fig. 1). Open pit mining was employed to recover the manganese ore. Overburden, largely broken, massive, dense novaculite beds, 25 to 75 feet thick, was drilled by jack hammer, blasted and removed by bulldozer. A 25-foot face of ore was exposed for about 100 feet in length in the sidehill open cut. The width of the ore zone was about 50 feet. The mine workings are on the north flank of a steep-sided, east-west trending hill, and are about 500 feet above the base of a valley to the north. Psilomelane predominates, and lesser quantities of pyrolusite and manganite are present. The deposit appears to occur in the lowermost beds of the upper Arkansas Novaculite division. Quartz veins are present locally and in association with the manganese. About 100 feet east of this cut, manganese ore was mined from an irregular open cut trending east-west that is about 200 feet long, from 75 to 175 feet deep, and 50 to 75 feet wide. On the east and south faces of this cut, overburden thickness ranges from 25 to 75 feet. The novaculite beds that are mineralized strike east-west, and dip from 20° to 75° north. The exposed ore face extends from the top of the cut to the base, and averages about 50 feet in width. At the top of the cut, the manganese ore is high in iron content, as it contains visible quantities of goethite or limonite.

Mining equipment used in the mine operation included air compressors, jack hammers, bulldozers, and front-end loaders. Dump trucks were used to haul the ore to a concentrating mill.

TABLE 2. -- MANGANESE MINES AND PROSPECTS IN MONTGOMERY COUNTY

Location	Active period	Unit	Type of deposit	Production	
				Quantity	Grade
NW¼NW¼ sec. 13, T. 2 S., R. 26 W.	1958-1959	Hog Jaw mine	Replacement in shale with associated quartz veins; psilomelane and pyrolusite.	1,800 tons	38-45% (c)
SE¼NW¼ sec. 10, T. 4 S., R. 27 W.	1886-1958	North Mountain mine	Fracture fillings in novaculite; psilomelane.	150 tons	56% (c)
SW¼NW¼ sec. 13, T. 4 S., R. 24 W.	1915	Nelson mine	Veins, pockets, cavities; manganite and psilomelane.	58 tons	---
NE¼NE¼ sec. 18, T. 4 S., R. 24 W.	---	Jones Valley mine	Fracture filling in novaculite; manganite and psilomelane.	●	---
N½ secs. 28-29, T. 4 S., R. 26 W.	---	United Mineral Industries prospect	Fracture filling in novaculite; nodules of psilomelane.	---	27% (c)
SW¼SE¼ and NE¼SW¼ sec. 11, T. 4 S., R. 23 W.	---	Plemmons-Woodall mine	Terrace gravels; cemented with manganite, psilomelane and wad.	---	---
E¼ sec. 18, T. 4 S., R. 26 W.	---	Willie Cogburn prospect	Fracture filling in novaculite.	---	▲
SE¼ sec. 14, T. 4 S., R. 26 W.	---	Monroe-Knold	---	---	---
E½ sec. 16, T. 4 S., R. 26 W.	---	Montgomery Manganese Corp.	---	---	---
Sec. 13, T. 4 S., R. 27 W., and sec. 18, T. 4 S., R. 26 W.	---	Polk Creek Mountain prospects	---	---	---
Secs. 17-18, T. 4 S., R. 27 W.	---	Sugartree Mountain	---	---	---

●No production reported.

▲scant milling grade ore.

(c) concentrate

TABLE 3. -- MANGANESE MINES AND PROSPECTS IN PIKE COUNTY

Location	Active period	Unit	Type of deposit	Production	
				Quantity	Grade
SW¼SW¼NW¼ sec. 18, T. 5 S., R. 23 W.	---	Frank Brunson prospect	Sparse manganese oxide with iron oxide.	---	---
S½SE¼ sec. 1, T. 5 S., R. 24 W.	1916	Watkins-White prospect	Seams and veins of wad and psilomelane with limonite.	---	---
SW¼NW¼ sec. 6, T. 5 S., R. 24 W.	---	Reynolds Mountain prospect	Veins, psilomelane with manganite, brown iron oxide.	---	---
NE¼SE¼, NE¼SW¼, SW¼NE¼, SE¼NE¼ sec. 1, T. 5 S., R. 25 W.	---	Kettelberger prospect, formerly Reynolds Mountain	Pyrolusite and manganite.	---	---
Center S½N½ sec. 9, T. 5 S., R. 25 W.	1916	Fagan mine	Thin veins of sparse manganite with iron oxide.	---	---
NE¼NW¼ sec. 8, T. 5 S., R. 25 W.	---	Bear Mountain prospect	Psilomelane and manganite lenses along bedding.	---	---
SW corner sec. 5, NW corner sec. 8, T. 5 S., R. 25 W.	1916;	Brushy Mountain mine	Stalactitic psilomelane and cement or veins in breccia.	1,000 tons?	12% (o)
NW¼SE¼ sec. 11, T. 5 S., R. 26 W.	1916	N. F. White prospect	Thin seams of psilomelane and manganite.	---	---
NW¼SW¼NW¼ sec. 10, T. 5 S., R. 26 W.	1916	Rattlesnake prospect	Few limonite and psilomelane veins. Maximum width 9 inches.	---	---
S½NE¼ sec. 6, T. 5 S., R. 26 W.	1916; 1954	Still mine (Sausman adit)	Mostly wad, some psilomelane with disseminated manganite. Partly in clay. Partly veins.	45 long tons	45% (c)

(o) ore  
(c) concentrate

— Continued on next page —

TABLE 3. -- MANGANESE MINES AND PROSPECTS IN PIKE COUNTY -- Continued

Location	Active period	Unit	Type of deposit	Production	
				Quantity	Grade
SE $\frac{1}{4}$ SE $\frac{1}{4}$ NE $\frac{1}{4}$ sec. 6, T. 5 S., R. 26 W.	1916	Prospect No. 3	Psilomelane with minor wad and limonite in clay, zone 7 feet wide.	---	---
SW $\frac{1}{4}$ SW $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 5, T. 5 S., R. 26 W.	1916	Prospect No. 2	Small pockets and veins of wad mixed with clay. Total is very small part of the whole.	---	---
SW $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 5, T. 5 S., R. 26 W.	1950's	Hog Pen Mountain	Fracture filling, psilom- elane and pyrolusite.	100 tons	46% (c)
SE $\frac{1}{4}$ SE $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 6, T. 5 S., R. 26 W.	1916	Risner Mountain prospects	Very thin veins of psilom- elane, limonite, and ferruginous manganese oxide.	---	---
NE $\frac{1}{4}$ SW $\frac{1}{4}$ NE $\frac{1}{4}$ sec. 3, T. 5 S., R. 26 W.	---	R. M. Cogburn prospect	Boulders of novaculite display small veins and pockets of psilomelane.	---	---
SW $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 6, T. 5 S., R. 24 W.	1917	Featherstone mine	Psilomelane and limonite in fractures and nodular manganese in clay.	40 tons	---
SW $\frac{1}{4}$ SW $\frac{1}{4}$ sec. 5, S $\frac{1}{2}$ SE $\frac{1}{4}$ sec. 6, T. 5 S., R. 25 W.	1950's	North American Manganese Co.	Pockets of psilomelane with inclusions of pyrolusite in novaculite.	---	---
NW $\frac{1}{4}$ NE $\frac{1}{4}$ , E $\frac{1}{2}$ NE $\frac{1}{4}$ sec. 8, W $\frac{1}{2}$ NW $\frac{1}{4}$ , SE $\frac{1}{4}$ NW $\frac{1}{4}$ , S $\frac{1}{2}$ NE $\frac{1}{4}$ , sec. 9, T. 5 S., R. 26 W.	---	Joseph Kehr prospect	Psilomelane in breccia zone and narrow veins in faults.	---	---

(c) concentrate

TABLE 4. -- MANGANESE MINES AND PROSPECTS IN POLK COUNTY

Location	Active period	Unit	Type of deposit	Production	
				Quantity	Grade
<b>COON CREEK AREA</b>					
SE $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 13, T. 4 S., R. 28 W.	1942, 1956-1959	Coon Creek mine	Psilomelane in fractured nova- culite. Some replacement.	1540 tons	45% (c)
<b>MISSOURI MOUNTAIN AREA</b>					
SE $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 36, T. 3 S., R. 28 W.	1956-1959	Lost Hatchet mine	Psilomelane in novaculite.	3665 <sup>*</sup> tons	44% (c)
<b>BOAR TUSK MOUNTAIN-SUGARTREE MOUNTAIN--EAGLE MOUNTAIN AREA</b>					
Center W $\frac{1}{2}$ SW $\frac{1}{4}$ sec. 27, T. 3 S., R. 30 W.	1888	Arkansas Development Co. mine (Ward mine)	Veins, pockets of pyrolusite and psilomelane.	3665 tons ore	---
NW $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 25, T. 3 S., R. 30 W.	1956-1959	Eagle Mountain mine	Lenticular pocket in frac- tured novaculite.	Small	4.0 % (o)
Parts of SE $\frac{1}{4}$ sec. 19, SW $\frac{1}{4}$ sec. 20 and NW $\frac{1}{4}$ sec. 29, T. 3 S., R. 30 W.	---	Boar Tusk prospect	Fractured novaculite.	---	7.64% (o)
Parts of SE $\frac{1}{4}$ sec. 27, N $\frac{1}{2}$ NE $\frac{1}{2}$ sec. 34, S $\frac{1}{2}$ S $\frac{1}{2}$ sec; 26, T. 3 S., R. 30 W.	---	American prospect	..... do. ....	---	5.5- 6.34% (o)

<sup>\*</sup> Largest producer in Polk County.

(o) ore  
(c) concentrate

-- Continued on next page --

TABLE 4. -- MANGANESE MINES AND PROSPECTS IN POLK COUNTY -- Continued

Location	Active period	Unit	Type of deposit	Production	
				Quantity	Grade
SE $\frac{1}{4}$ SE $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 5, T. 4 S., R. 30 W.	---	West Hanna Mountain (Bowen claim)	Psilomelane and manganite seams to 3 inches thick and 3 feet long in fractured novaculite.	---	---
SW $\frac{1}{4}$ SW $\frac{1}{4}$ sec. 1, T. 4 S., R. 30 W.	---	West Hanna Mountain (W. Allen claim)	Psilomelane and manganite seams to 3 inches thick in fractured lower novaculite.	---	---
NE $\frac{1}{4}$ NE $\frac{1}{4}$ NE $\frac{1}{4}$ sec. 10, T. 4 S., R. 30 W.	---	Prospect	Manganese ore in veins and pockets in novaculite.	---	37.96% (c)
SW $\frac{1}{4}$ NW $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 10, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	10.35% (o)
Center N $\frac{1}{2}$ N $\frac{1}{2}$ sec. 10, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	3.48% (o)
SE $\frac{1}{4}$ NE $\frac{1}{4}$ SW $\frac{1}{4}$ sec. 10, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	6.17% (o)
SE $\frac{1}{4}$ NW $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 10, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	2.12% (o)
SE $\frac{1}{4}$ SW $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 5, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	0.29% (o)
SE $\frac{1}{4}$ SW $\frac{1}{4}$ SE $\frac{1}{4}$ sec. 5, T. 4 S., R. 30 W.	---	.....do.....	.....do.....	---	15.32% (o)
<b>HATTON-BOG SPRINGS AREA</b>					
NW $\frac{1}{4}$ sec. 10, T. 5 S., R. 32 W.	---	Lavine prospect	Manganese stains and psilom- elane lenses in novaculite.	---	---
NW $\frac{1}{4}$ and NE $\frac{1}{4}$ sec. 14, T. 5 S., R. 32 W.	---	Hablitzel prospect	Psilomelane in novaculite.	---	5-10% (o)
NW $\frac{1}{4}$ sec. 18, T. 5 S., R. 32 W.	---	Doty prospect	---	---	---
<b>MISCELLANEOUS</b>					
SW $\frac{1}{4}$ NE $\frac{1}{4}$ sec. 6, T. 4 S., R. 29 W.	1921; 1942; 1952	Sugarstick mine	Ore on bedding planes in novaculite.	18 carloads	40 <sup>+</sup> % (c)
NE $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 33, T. 3 S., R. 29 W.	1951	Dr. Scott claim and prospect	---	22 tons	40 <sup>+</sup> % (c)
SE $\frac{1}{4}$ NW $\frac{1}{4}$ sec. 7, T. 4 S., R. 29 W.	---	Pickel Manganese prospect	Massive to concretionary psilomelane and pyrolusite, 4- and 6-foot zones in novaculite.	---	---
NW $\frac{1}{4}$ sec. 29, T. 4 S., R. 28 W.	---	E. S. White prospect	Psilomelane in fractures and replacing novaculite.	---	5-10% (o)

(o) ore

(c) concentrate

### Coon Creek Mine

This mine in the SE $\frac{1}{4}$ NW $\frac{1}{4}$  sec. 13, T. 4 S., R. 28 W., Polk County (Fig. 1) is in the vicinity of the W. A. Davis prospect No. 2 first described by Miser (8). Most recent activities were in the fall of 1958. Mine workings, consisting of two sidehill open cuts and several shallow pits, are on the north side near the base and on the flank of an east-west trending hill. The cuts face on a small intermittent stream tributary to Long Creek. The open cut near the base of the hill is about 75 feet long and 25 feet wide. A mineralized zone 10 to 20 feet thick is exposed for about 60 feet in length. The massive and dense novaculite host rock strikes east-west and dips 20° to 25° south. The ore occurs at or near a fault zone. In an open cut, higher in elevation than the first described mine working, manganese and iron minerals occur in significant amounts. The latter pit is about the same size as the lower workings. This mine was worked in much the same manner as the Lost Hatchet mine. The deposit also contains a suite of iron phosphate minerals, turquoise (?), and possibly chrysocolla.

### Hog Jaw Mine

The Hog Jaw mine was first opened in 1958 in the NW $\frac{1}{4}$  NW $\frac{1}{4}$  sec. 13, T. 2 S., R. 26 W., Montgomery County. Mine workings are situated on the south flank of a low hill underlain by an upper shale sequence in the Arkansas Novaculite, and face an intermittent stream on the southwest. Manganese ore is associated with quartz veins that fill fractured zones in the shale, and in part, the manganese replaces the shale. Goethite, pyrite, and thin veined barite are abundant locally. Overburden and mining cuts were made by bulldozer in an area 200 feet north to south and 400 feet east to west. The manganese and quartz veins are concentrated at joint intersections in the shale. The manganese ore was drilled by jack hammer, blasted, and removed by hand mining. Some exploratory drilling was done in an effort to find ore extensions, without success.

### Eagle Mountain, Hog Pen Mountain, and North Mountain Mines

In 1958, a small production of manganese was recovered from three mines in areas formerly described in part by Miser (8). The Eagle Mountain mine is in the NW $\frac{1}{4}$ SE $\frac{1}{4}$  sec. 25, T. 3 S., R. 30 W., Polk County (Fig. 1). At this location, manganese ore of generally poor quality was hand mined from several symmetrical pockets, about 8 feet wide and 50 feet long. The host beds of massive, dense novaculite strike east-west and have generally steep dips, some near vertical. Folding is notable, and a small anticline is exposed in an open cut on a northward facing slope. The manganese occurs in small fractures and as pods and has a substantial associated iron content.

Hog Pen Mountain mine is in the SW $\frac{1}{4}$ NW $\frac{1}{4}$  sec. 5, T. 5 S., R. 26 W., Pike County (Fig. 1). Manganese minerals fill fractures in a vertical dipping, broken zone of hard, dense novaculite. The mineralized area, as exposed,

is about 6 feet wide and 500 feet long. The ore was mined through an inclined adit that in part penetrates the ore zone. Reportedly, one carload lot of ore from this property assayed 44.15 percent manganese, 0.34 percent phosphorus and 12 percent silica.

North Mountain mine is in the SE $\frac{1}{4}$ NW $\frac{1}{4}$  sec. 10, T. 4 S., R. 27 W., Montgomery County. The mine was first active in about 1886, and the deposit was last worked in 1958. Manganese occurs as fillings in a heavily fractured east-west trending band of novaculite that is overturned to the north at a steep angle. Much of the better grade of ore appears to be in both the upper and middle divisions of the Arkansas Novaculite. Cuprite, native copper, malachite, turquoise (?), and possibly chrysocolla are associated with the manganese as are iron minerals and very thin quartz veins. The exact relationship of copper-bearing silicified zones associated with the manganese deposit is not known partly because mine tunnels are virtually inaccessible. The copper minerals are found sparingly in dump material apparently removed from a tunnel working.

In one place, a bulldozer cut 25 feet wide and 100 feet long crosscuts the ore body. On the west side of the cut an adit was driven adjacent and along the strike of the ore body. The ore as mined was hauled to a mill for concentration.

### Mills

In the early days of mining in the West-Central District, manganese was recovered by simple hand mining methods, and ore was hand cobbled for direct shipment to markets. Small mills for ore concentration were built during World Wars I and II using jigging to recover shipping-grade ore. Variations of log washers were also used to obtain ore concentrates but were largely unsuccessful in that poor recoveries were made. During the decade of the 1950's, when larger quantities of mill grade ores were made available through use of mechanized mining equipment such as bulldozers, larger concentrating plants were operated. Some milling operations were moderately successful in treating selected ores, and concentrates did not contain substantial quantities of iron, silica, or phosphorus.

The Federal Bureau of Mines in the mid-1950's conducted beneficiation tests on ores from several properties in the West-Central Manganese District (6). Some ores were amenable to concentrating methods employed, others were not because of the retention of fine grained impurities that made the concentrate unacceptable with relation to specification requirements.

At least two mills that operated in the late 1950's produced marketable concentrate. These are described in subsequent paragraphs.

The largest mill (Duncan, Dieckman and Duncan) known to be successful was operated in the NW $\frac{1}{4}$ SW $\frac{1}{4}$  sec. 33, T. 3 S., R. 29 W., Polk County (Fig. 1). This mill treated ore mined in bulk, from at least seven widely

separated mines in the district. The mill was equipped with a coarse ore bin with 5-inch grizzly; washing and screening trommel with 1-inch openings; picking belts; 24-inch jaw crusher for plus 1-inch material; bucket elevator; double trommel screen, one having 9/16-inch openings, the other 5/16-inch openings; the plus 9/16-inch size was passed through a roll crusher. Crushed material was passed through a 4- by 12-foot trommel having four screen sizes; minus 1/16- to 1/16-inch sizes went to a Wilfrey table for concentration of fines; plus 1/16- to 1/8-inch passed to a three-cell jig; plus 1/8- to 3/16-inch sizes went to a five-cell jig; plus 3/16- to 5/16-inch material passed to a five-cell jig; and, plus 5/16- to 9/16-inch material passed to a four-cell jig. Three products resulted—concentrates, middlings, and tailings. Periodically, middlings were rerun. Some tailings, which contained a high concentration of fine novaculite, was marketed as road gravel. Diesel-fueled engines furnished all power.

Water was supplied from a small stream pond using an 8-inch pump. The mill had a capacity of 200 tons per day. Manganese recovery was dependent on grade of mill feed, and usually a recovery of 70 to 80 percent of the contained manganese was possible.

At the time of examination by Bureau of Mines personnel in 1959, samples of the head feed, concentrates, middlings, and tailings were taken for analyses. The head feed at that time was run-of-mine ore from the Lost Hatchet mine. Results of analyses are as follows:

	Manganese in percent
Head feed-----	6.77
Concentrates-----	49.97
Middlings-----	15.22
Tailings-----	1.08

Another mill (Coon Creek Manganese Co.) successful at producing marketable concentrates was in the NE $\frac{1}{4}$ SE $\frac{1}{4}$  sec. 13, T. 4 S., R. 28 W., Polk County (Fig. 1). This mill patterned after the first described mill was equipped with washer and screen trommels, jaw crushers, roll crusher, bucket elevator, and two five-cell jigs and one three-cell jig. Reportedly, a satisfactory recovery rate was maintained.

Other smaller mills were in the SW $\frac{1}{4}$ NE $\frac{1}{4}$  sec. 5, T. 5 S., R. 26 W., Pike County (Hog Pen Mountain, Fig. 1); the

NW $\frac{1}{4}$ SE $\frac{1}{4}$  sec. 35, T. 3 S., R. 29 W., and the SW $\frac{1}{4}$ NW $\frac{1}{4}$  sec. 13, T. 4 S., R. 26 W., Montgomery County (Fig. 1); and, the NW $\frac{1}{4}$ NW $\frac{1}{4}$  sec. 32, T. 3 S., R. 29 W., in Polk County (Fig. 1).

## RESOURCES AND POTENTIALS

Manganese resources in the West-Central District have been tentatively evaluated as large because of insufficient exploration. In some places substantial mining has occurred which permits the gathering of measurable data that suggest that a few to several tens of thousands of tons of milling-grade ore would be available. The largest of these deposits may contain 40,000 to 50,000 tons. However, small deposits of relatively high-grade manganese material are scattered over a relatively large area and, if exploration was done, doubtlessly several deposits might yield sizable tonnages. In the aggregate, the tonnage potential in the district is significant, probably amounting to several hundred thousand tons containing 5- to 10-percent manganese. In most instances, little or no effort has been made to determine the horizontal extent or depth of any deposit. At the most productive mines, now idle, much low-grade manganese ore remains, but only in times of high prices and suitable markets is bulk mining warranted. Small quantities of high-grade ore could be hand mined at a number of sites, but manganese ore prices would have to be high to warrant mining. The association in some instances of other metallic minerals that might be recoverable, such as copper, could enhance the outlook for the district.

## PRODUCTION PROBLEMS

Exploration for manganese ore deposits would be expensive because of the character of the host rock (mostly massive, dense novaculite which is extremely hard and abrasive), the scattered nature of the known deposits, the rugged terrain, and the remoteness of the areas. Building of access roads would be difficult and expensive. Many prospects have been abandoned at shallow depths owing to high operating costs rather than exhaustion of ore minerals.

Mill-grade ore could be recovered at several sites but would have to be transported to a centrally located mill. Milling costs would be high, but improved milling techniques that yield higher recovery would decrease costs. Blending of ores to provide a uniform mill feed could improve milling costs and result in lower contents of silica, iron, and phosphorus. However, prior to research for other more efficient milling techniques, much exploration at probable high costs would be a prerequisite to determine the possible extent, character, and minability of the manganese ores.

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# CONTROLS ON THE DISTRIBUTION OF BARITE DEPOSITS IN ARKANSAS

By Jeffrey S. Hanor<sup>1</sup> and Lawrence R. Baria<sup>2</sup>

## ABSTRACT

The following model is proposed for the origin of the bedded barite deposits in the Ouachitas: at the beginning of Stanley time (Mississippian, late Meramecian), minor deformation and a change in tectonic regime along the Ouachita continental margin resulted in the increased influx of clastic materials and the formation of local submarine depressions near the base of the continental rise and slope. Fractures tapped normal subsurface formation waters, which contained dissolved barium. These brines migrated upward in response to a difference in hydrologic head and ponded in the local submarine basins. Barite precipitated at the interface between the brines and the overlying, sulfate-rich seawater and then sank downward to accumulate as massive beds. Additional Ba may have been removed from solution by adsorption on clay minerals settling through the brines. As the supply of subsurface brine was exhausted, barite deposition ceased.

Some  $2.5 \times 10^8$  years later, additional barite was deposited as cements in permeable Cretaceous sands and gravels at the present southern margin of the Ouachitas. The mobilization of these ore-forming solutions may have occurred in response to thermal gradients produced during the Late Cretaceous episode of intrusive activity in Arkansas.

It is felt that the models proposed in this paper can account for many features of the temporal and spatial distribution of known barite deposits in Arkansas and may possibly be of use in developing guides for finding new deposits.

## INTRODUCTION

The barite deposits of the Ouachita region of Arkansas are of considerable geochemical and economic interest. Over a 25 by 150 km area of central and west central Arkansas (Figure 1), Paleozoic siltstones and shales of the lower Stanley Group locally contain beds consisting of up to 85 weight percent  $\text{BaSO}_4$ . Of these deposits, the barite ore body of the Chamberlain Creek syncline, in the eastern Ouachitas, is one of the largest known single concentrations of barium in the earth's crust. For many years, the Baroid open pit and adjacent Magcobar-Dresser mine, both of which are developed in this deposit, were the largest producers of barite in the world. Additional barite deposits of younger age occur in Cretaceous sediments at the southern margin of the Ouachita region (Figure 1).

The primary purpose of this paper is to describe a model for the origin of the Paleozoic bedded barite deposits of the Ouachitas which can satisfactorily account for the time of deposition, sources of barium and of sulfate, and the unique set of geologic conditions which caused barite to be concentrated in economically valuable deposits. In addition, some comments will be made regarding probable controls on the distribution of the barite deposits in Cretaceous host rocks. It is hoped that some of the thoughts presented in this brief paper will be of use in the search for additional barite deposits.

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## PALEOZOIC BEDDED BARITE DEPOSITS

### Description

A detailed description of the bedded barite deposits is given by Scull (1958) whose discussion of the occurrence of barite in Arkansas remains the most comprehensive to date. Zimmerman's more recent work (1965) also contains a wealth of valuable information on the bedded deposits. We will summarize briefly some of the principal features of the bedded deposits.

Barite in the Ouachitas proper occurs as conformable, stratiform lenses and disseminations in shales and siltstones of the lowermost Stanley Group of Mississippian age. Most of the Stanley was deposited during Chesterian time. The lowermost Stanley, however, is probably late Meramecian in age (Gordon and Stone, 1973). The district-wide stratigraphic control of barite is remarkable. Recent field mapping (C. G. Stone and B. R. Haley, personal communication, 1973) has confirmed that most of the known bedded deposits occur in the lower Stanley, including deposits in western Arkansas originally thought to be in Devonian host rocks.

Individual deposits are lenticular and range in maximum stratigraphic thickness from less than one meter to 30 meters and length in outcrop ranges from 40 to over 7500 meters. The deposits are separated stratigraphically from the underlying Arkansas Novaculite by one to seven meters of unmineralized shale. The barite content exceeds 85 weight percent in some portions of the Chamberlain Creek deposit (McElwaine, 1946), but in most other deposits the maximum concentrations are less.

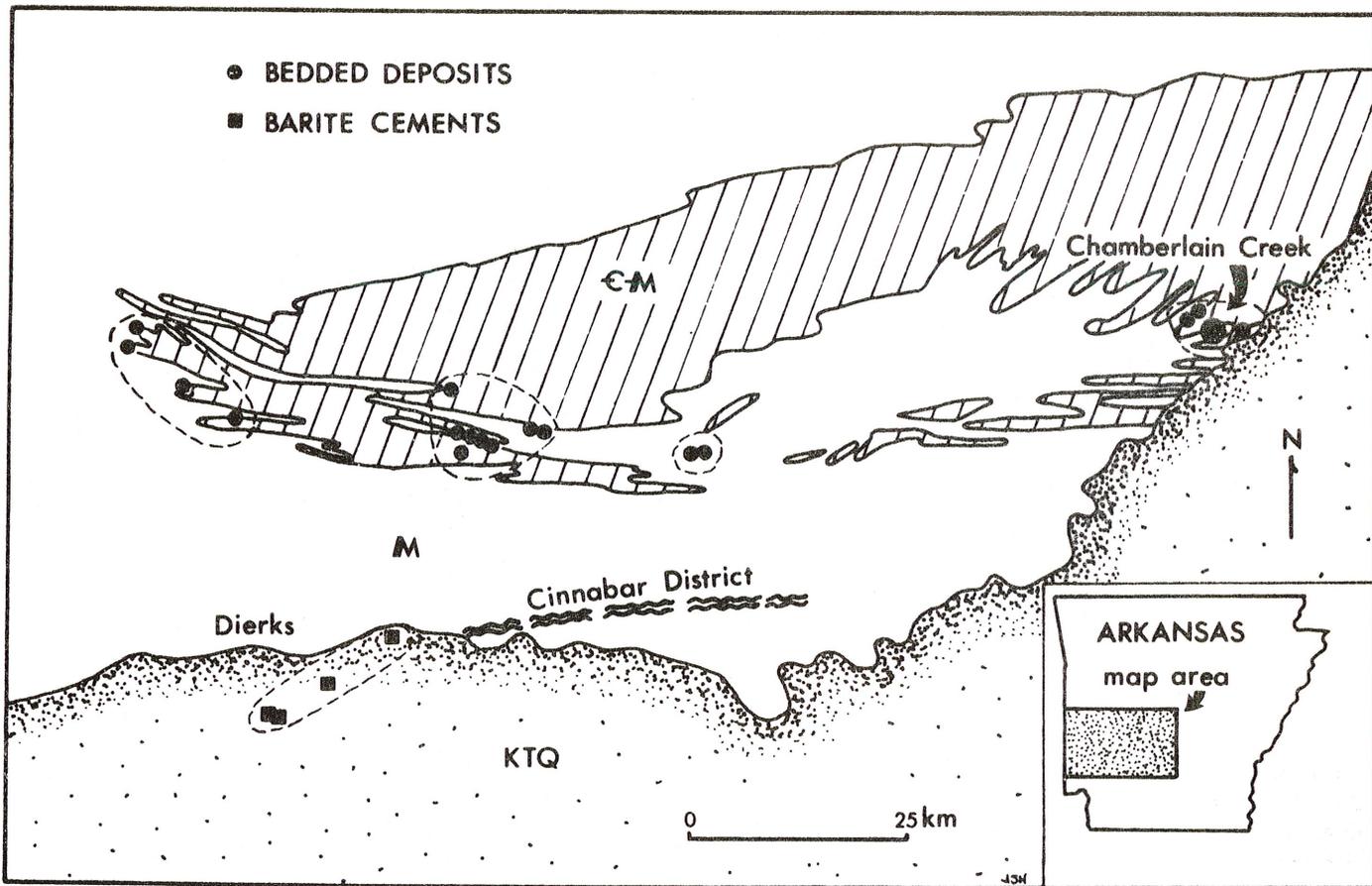


Figure 1. -- Sketch map of the Ouachita region of Arkansas, showing location of known bedded barite deposits (circles) and younger barite cement (squares). Locations from Scull (1958). C-M represents the outcrop area of pre-Stanley sediments, KTQ the area of outcrop of Cretaceous and younger sediments.

Within individual deposits, barite occurs principally as (1) disseminated single crystals ( $< 150 \mu\text{m}$  length) in a shale or carbonate matrix, (2) dense, finely crystalline laminae, and (3) nodules of spherically radiating crystals.

#### Time of Deposition

On the basis of the spatial proximity of the Magnet Cove alkali intrusive complex (Erickson and Blade, 1963) to the Chamberlain Creek deposit, Park and Branner (1932) originally postulated that the barite deposit was a hydrothermal replacement related to the nearby igneous intrusive activity. Scull (1958) expanded the concept of hydrothermal replacement to include all of the Ouachita deposits. He suggested that the deposits were Cretaceous in age, a time of extensive igneous activity in Arkansas. Beginning in 1961, Zimmerman and Amstutz (1961, 1964a, 1964b) presented a series of arguments based on pioneering textural and sedimentological studies that convincingly demonstrated that the barite was deposited penecontemporaneously with host clays and silts. This information is discussed in detail by Zimmerman (1965). Further independent evidence corroborating a sedimentary origin has been developed by Brobst and his colleagues (Brobst and Ward, 1965; Miller and others, 1972; and Brobst, 1973), by Hanor (1966), and by Baria (in preparation).

We believe that the barium now observed in the deposits was emplaced at the time of deposition of the host sediment. In this sense, then, the deposits are sedimentary in origin. The precipitation of the mineral barite probably occurred in a number of different ways over a considerable period of time. Most of the barite, however, was probably precipitated during the depositional and early diagenetic history of the sediment. Locally, there are textures suggesting primary precipitation from sea water, early replacement of other minerals by barite, penecontemporaneous mechanical transport and redeposition of barite, and later recrystallization of barite. Many of these diagenetic features have been described by Zimmerman (1965), and others are currently under investigation (Baria and Hanor, 1974).

The lithologic break between the Arkansas Novaculite and the overlying Stanley Group marks a profound discontinuity in the sedimentological and tectonic development of the Ouachitas. At the inception of Stanley time, when the barite was deposited, sedimentation rates increased abruptly, probably in response to a change in tectonic regime which permitted the productive waters and siliceous zones of the Ouachita basin (Lowe, 1973) to be flooded by terrigenous debris. This change in tectonic style is marked also by the deposition of the Hatton tuffs,

which lie approximately 100 m above the barite at the Chamberlain Creek deposit.

#### Source of Barium

In looking for possible sources of Ba for the Stanley deposits it is pertinent to consider sites of more recent deposition of sedimentary barite (Hanor and Baria, 1973). For example, barite is a very widespread and abundant authigenic mineral in Cenozoic pelagic sediments of the Pacific. Some surface sediments of the equatorial Pacific and East Pacific Rise contain over 5 weight percent barite (Arrhenius and Bonatti, 1965; Hanor and Brass, 1968; Boström *and others*, 1973). Recent calculations of Ba accumulation rates in the Pacific (Hanor, 1972; Boström *and others*, 1973), however, have shown that maximum pelagic Ba fluxes are less than 10 mg Ba/cm<sup>2</sup>/10<sup>3</sup> y. At maximum pelagic rates of Ba supply, it would have required approximately 500 million years to have produced a Chamberlain Creek deposit, 30 m in thickness and containing, on the average, 50 weight percent barite. Because of the apparent absence of pelagic processes efficient enough to produce high Ba fluxes at the seafloor, it seems necessary to invoke a submarine source for Ba. For the reasons developed below, we believe that normal subsurface formation waters would be adequate.

When marine waters became trapped and buried in sediments containing organic matter, bacterial reduction removes sulfate from solution and permits processes of dissolution to increase the Ba concentration. This evolution in water composition is illustrated by the example shown in Table 1. Column A is the average composition of Pliocene and Miocene marine formation waters in the Wilmington, California oil fields (Gates and Caraway, 1965). Column B shows an analysis of normal seawater of the same chloride content. Column C shows the ratio (concentration in formation water/concentration in seawater) for each constituent.

While there is some evolution in major cation composition, by far the most profound effects are the reduction of sulfate and the substantial increase in Ba. The high Ba levels in these very waters have caused serious problems when oil field operators have injected seawater into the formations to drive out additional oil (Gates and Caraway, 1965). The mixing of high-Ba formation water and seawater has resulted in the precipitation of a barium sulfate scale in producing wells.

A summary of the Ba content of subsurface formation waters has been presented by Hanor (1966) and Puchelt (1967, 1972). Puchelt lists waters containing as much as 5500 ppm Ba. The average Ba content of Na-Cl and Na-Ca-Cl subsurface waters included in the compilation of White *and others* (1963) is 200 ppm, with a high value of 1550 ppm. It seems reasonable that subsurface waters of this general type formed in the lower Paleozoic sediments of the Ouachita basin and were present at the time of the tectonic change that marked the beginning of Stanley sedimentation. As discussed below, fractures and minor faults produced at this time could have served as conduits for the upward injection of brines into bottom sediments and waters. The brines mixed with sulfate-rich normal marine waters and barite was precipitated (see Hanor, 1969).

As an example of the amount of brine which may have been involved, we estimate (Hanor and Baria, in preparation) that 1.6 x 10<sup>14</sup> liters of a normal formation water containing 200 ppm Ba would have been necessary to form a deposit the size of the Chamberlain Creek and adjoining occurrences. This quantity of fluid could be contained in a volume of subsurface sand with a porosity of 15%, a thickness of 300 m (the approximate thickness of the underlying Blaylock Sandstone in the vicinity of Chamberlain Creek), and a radius of 33 km.

TABLE 1. -- COMPARISON OF CALIFORNIA PLIOCENE - MIOCENE MARINE FORMATION WATERS (GATES AND CARAWAY, 1965) AND SEAWATER OF THE SAME CHLORIDE CONTENT, SHOWING HOW DISSOLVED BARIUM INCREASES AND DISSOLVED SULFATE DECREASES WITH BURIAL.

Species	A Formation Waters mg/l	B Seawater mg/l	C Ratio of (A/B)
Na	10,485	10,301	1.02
K	297	370	0.80
Mg	362	1,239	0.29
Ca	507	395	1.28
Ba	68	0.01	6800.00
Cl	18,528	18,528	1.00
HCO <sub>3</sub>	916	136	6.74
SO <sub>4</sub>	4	2,596	0.0015

## Source of Sulfate

If the Ouachita barite was precipitated by mixing a Ba-rich brine with seawater, the sulfur and oxygen isotopic composition of the barite should be close to that of seawater sulfate. A barite sample collected by Hanor in 1963 from the Chamberlain Creek deposit was analyzed by T. M. Church. The results are  $\delta O^{18}$  (SMOW) = 13.2 ‰;  $\delta S^{34}$  (Troilite) = 18.3 ‰. These values are consistent with seawater sulfate. In fact, the  $\delta S^{34}$  value corresponds closely to the suggested sulfur isotopic composition of middle Mississippian seawater (Thode and Monster, 1965; Holser and Kaplan, 1966).

## Mechanism of Concentration

Regional stratigraphic and sedimentological studies of the later Paleozoic sediments of the Ouachitas would tend to place the sites of barite deposition near the base of the Mississippian continental slope, perhaps at depths exceeding 2000 m (Figure 2).

The tectonically active nature of the Ouachita trough at this time would have favored the formation of local submarine depressions (Guilcher, 1963). Dense, submarine brines would pond in these depressions, and much of the precipitation of barite would thus tend to be confined to relatively localized areas. In the absence of such a mechanism for concentration, simple injection of a brine into seawater would produce a highly dispersed and economically unimportant precipitate.

Removal of Ba from a brine pooled in a depression on the seafloor could be accomplished in one of two ways (Figure 3): First, precipitation of  $BaSO_4$  would occur at the brine-seawater interface. The barite would sink through the brine and accumulate as massive beds. Second, downward transport of Ba could occur by adsorption on clay particles. As one variant of this latter mechanism there exists the intriguing possibility that turbidity currents would flow out across the top of a dense brine. Clays, originally in equilibrium with seawater, would adsorb Ba from the brine. Given a later source of sulfate, this Ba could be later stripped off the clays, possibly to form nodular or concretionary beds of barite.

We have tested the mechanical feasibility of the first mode of precipitation by setting up a stratified column, where a  $Na_2SO_4$  solution overlay a denser  $NaCl-BaCl_2$  solution. Arborescent  $BaSO_4$  crystals formed at the interface between the two solutions, sank, and accumulated in a mass at the bottom of the column. The unusual habit of the crystals (Figure 4a) is similar to some barite we have recently found at Chamberlain Creek (Figure 4b). It is possible that the crystals in Figure 4b also formed at a solution interface.

## Why did Deposition Cease?

Although the Ouachita trough remained tectonically active throughout the upper Mississippian and

Pennsylvanian, scant bedded barite deposits have been found higher in the Stanley or in later Paleozoic sediments. Deposition of economic concentrations of barite could have ceased for one of several reasons:

1. The submarine topographic depressions were filled up and no longer acted as sedimentary traps.
2. An increase in rates of clastic sedimentation diluted the barite to low concentration levels.
3. The supply of Ba into the system stopped.

Recent work by Baria (in preparation) on the distribution of the Hatton Tuff in the eastern Ouachitas indicates that the tuff beds are thickest in vicinity of the Chamberlain Creek deposit. It is possible, therefore, that the area was still a basin at the time of tuff deposition and that the first reason given above is not valid. Above the barite zone, the Stanley contains many more sand units, and it is likely that an increase in the flux of clastic material was significant in diluting Ba levels. The overall controlling factor, however, may have been simply that the supplies of Ba were exhausted. Subsurface reservoirs are finite, and there was simply not enough volume of fluid to continue supplying Ba at the fluxes necessary to form significant concentrations.

## DEPOSITS IN CRETACEOUS SEDIMENTS

At the southern margin of the Ouachita Region in the vicinity of the town of Dierks (Figure 1) isolated deposits of barite occur as cements in sands and gravels of the Trinity Formation of Cretaceous age (Scull, 1958). At least one of the deposits has been mined briefly for gravel and barite (McElwaine, personal communication, 1963).

It is significant to us that the deposits lie approximately along an extension of the trend of the Arkansas cinnabar deposits which are developed in Paleozoic host rocks to the east (Reed and Wells, 1938). Minor barite is known from the westernmost Hg deposits. We would suggest that the Ba and Hg are genetically related and that the apparent lateral separation of barite and cinnabar deposits may be a manifestation of vertical zonation. Ba may have been precipitated from fluids ascending into highly permeable Cretaceous sands and gravels, where there was active circulation of sulfate-bearing groundwaters. The Hg precipitated as a sulfide in the less permeable, more reducing environment below.

## RELATION TO IGNEOUS ACTIVITY

The relation between barite deposition and igneous activity in Arkansas has quite properly been a favorite topic for at least 40 years (see Stone and Sterling, 1964), and we would be remiss not to discuss it here. We see no need on the basis of present evidence to invoke an igneous source (by any definition) for the barite. There may, however, be more subtle, but still important

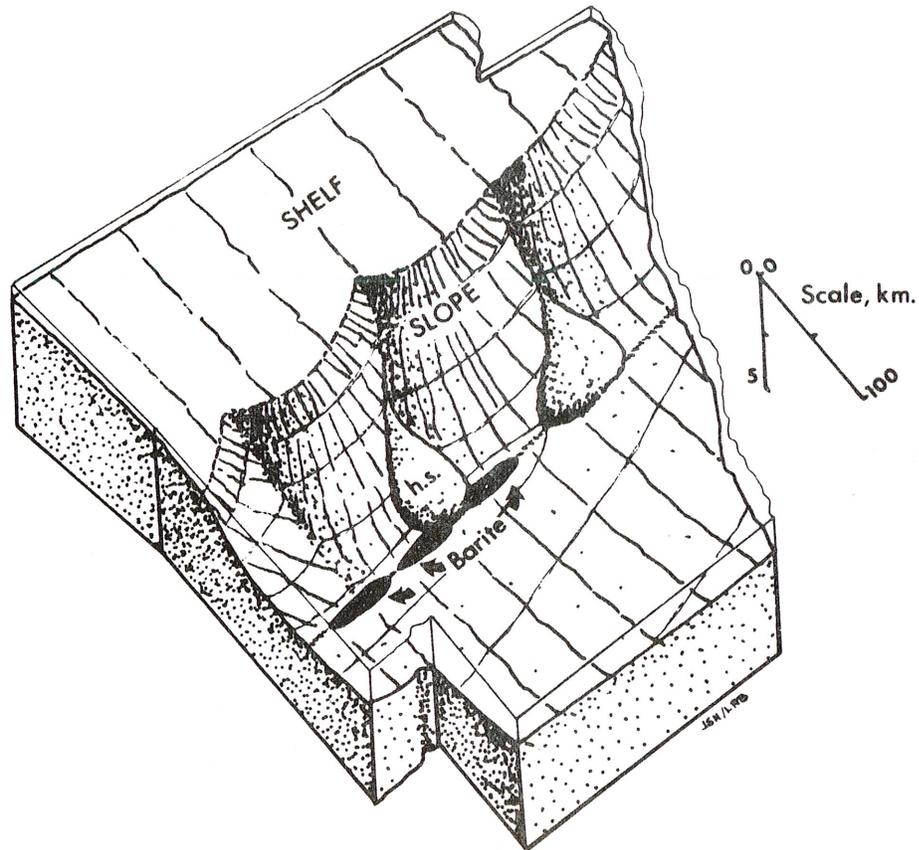


Figure 2. -- Possible configuration of Ouachita continental margin at lower Stanley time (late Meramecian). Ba-rich brines may have pooled in local depressions near the base of the continental slope. The vertical scale is highly approximate. Deposition of barite, however, may have occurred at depths in excess of 2000 m. The lobe marked "h.s." represents the Hot Springs Sandstone, a basal Stanley sand unit.

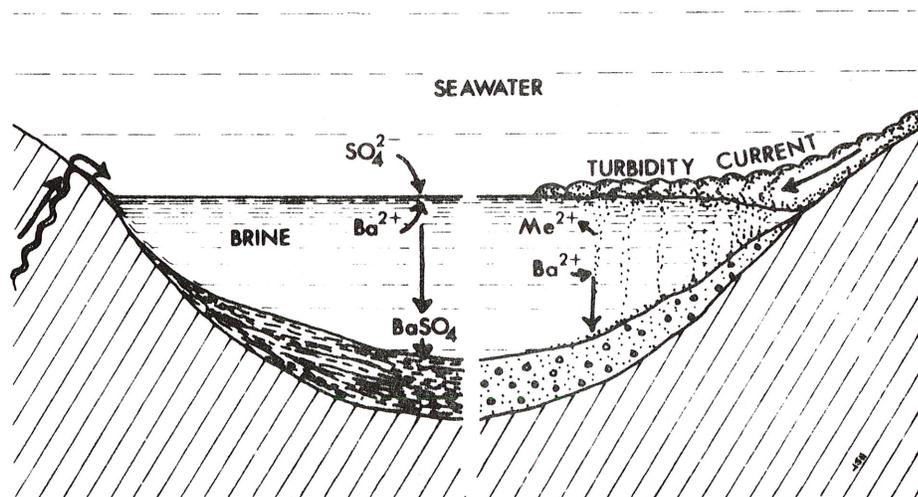


Figure 3. -- A cross-section of a hypothetical submarine brine pool. On the left, dense formation water debouch onto the sea floor and pond in a submarine depression. Dissolved Ba in the brine reacts with seawater sulfate to form barite, which then accumulates as massive beds. On the right, dissolved Ba is adsorbed by clays settling through the brine. Ba-rich clay beds may later give rise to nodular barite.

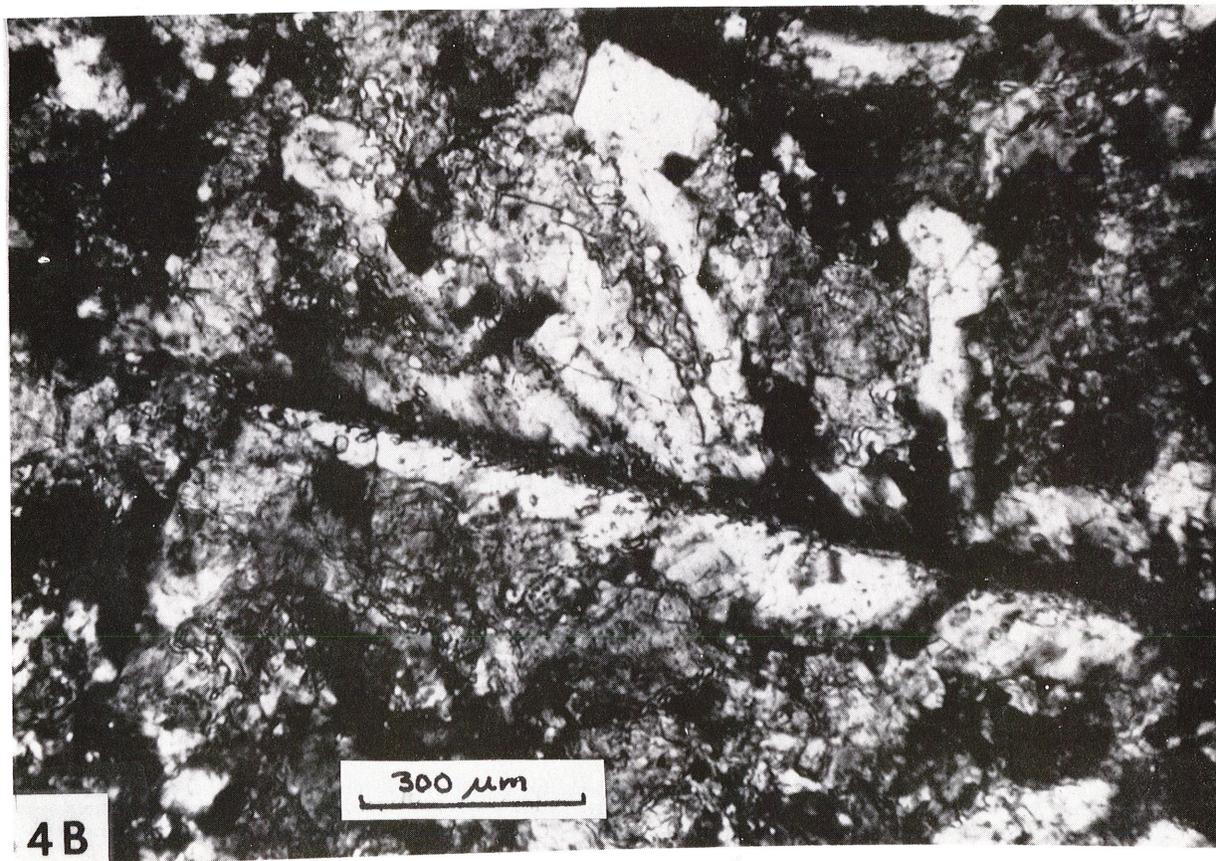


Figure 4. -- Photomicrographs of (A) arborescent  $\text{BaSO}_4$  crystals formed experimentally at a brine interface and (B) arborescent barite crystals found in the Chamberlain Creek deposit (cross polarized light). Both photos are at the same scale.

connections. The Hatton Tuff volcanism, which postdates barite deposition, may have been preceded by an increase in regional heat flow in the Ouachita trough. It has been postulated (Hanor, 1973) that changes in the heat flow regime of a sedimentary basin could trigger gravitational instability of subsurface brines. This would enhance the possibility of upward migration of sedimentary fluids in early Stanley time. Similarly, thermal gradients associated with Cretaceous igneous activity may have driven the fluids which supplied Ba for the Dierks district. Finally, it is possible that the location of the Magnet Cove intrusive was controlled in part by the same zone of crustal weakness that had resulted in the upward migration of brines some  $2.5 \times 10^8$  years earlier.

### APPLICATIONS

If our proposal is correct, and the bedded barite deposits of Arkansas formed from subsurface brines which pooled in depressions on the floor of the lower Stanley sea, then the following guides may be useful in locating new bedded deposits:

1. Through a detailed study of the paleogeography and depositional environment of the lower Stanley it might be possible to recognize a pattern to the distribution of local basins, the potential sites of barite accumulation.

2. Because volumes of subsurface brines are limited, there may be a maximum limit to the total amount of barite which could have been deposited within regions of equivalent size. For example, the low-grade and laterally extensive deposits of the western Arkansas districts may reflect a less effective ponding or localization of brines,

where an amount of barite perhaps equivalent to that in the Chamberlain Creek region was deposited over a much larger surface area. Such regions may not contain massive, high-grade deposits.

3. If we are correct and the source of Ba was formation water originally trapped in sandy units, the distribution of barite may in part reflect the distribution and thickness of underlying sand units, such as the Silurian Blaylock Sandstone, or the Ordovician Blakely and Crystal Mountain Sandstones.

Within the Dierks barite district, additional barite deposits may be found in permeable Cretaceous sediments along the barite-cinnabar trend. From an economic standpoint, however, such deposits may be less important for barite than for the possible Hg mineralization which may lie below.

### ACKNOWLEDGEMENTS

It is a pleasure to acknowledge the continued cooperation we have received from the staff and management of the Baroid Division, NL Industries, Inc.; the Magnet Cove Barium Corporation-Dresser Industries; and the Arkansas Geological Commission. Our work has been made possible in part by the previous contributions of B. J. Scull, R. A. Zimmerman, D. A. Brobst, and others on the barite deposits of Arkansas. We thank D. R. Lowe and R. D. LoPiccolo for many valuable discussions. Hanor is particularly indebted to the work of J. Francis Williams, whose classic treatise on the igneous rocks of Arkansas (1891) was instrumental in stimulating his interest in geology.

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# IRON PHOSPHATE MINERALS OF THE OUACHITA MOUNTAINS, ARKANSAS

By Albert L. Kidwell<sup>1</sup>

## ABSTRACT

A characteristic suite of iron phosphate minerals occurs in fracture zones of the Arkansas Novaculite along an east-west strike distance of almost 30 miles in southern Montgomery and Polk Counties, Arkansas. The mineral assemblage includes the ferrous-ferric phosphate minerals laubmannite, rockbridgeite, lipscombite, beraunite, dufrenite and a new mineral, and the ferric phosphate minerals strengite and cacoxenite. Turquoise is commonly associated as one of the last minerals to form. The minerals occur upon earlier goethite and an amorphous, red iron phosphate. Manganese oxides are common associates, but manganese is not an essential element in any of the phosphate minerals. The well-known Arkansas deposits of the aluminum phosphates wavellite and variscite generally do not contain any of the iron phosphates, and vice versa. One exception is at Wilson Springs where wavellite, variscite, strengite, and cacoxenite occur in the novaculite near the contact with an igneous intrusion. The source of the phosphate in all cases is believed to be the novaculite itself from which it was leached and redeposited as iron phosphates.

## INTRODUCTION

The writer was first introduced to the iron phosphates in the summer of 1948 while working on a survey of economic minerals in western Arkansas. One of the spots visited was a prospect pit that was being dug by Mr. Isom Avants of the Shady Community, about ten miles southeast of Mena, Arkansas. Mr. Avants was frank to admit that he did not know the identity of the canary yellow mineral in tiny acicular crystals and the small clusters of rose colored crystals associated with it. Neither did the writer at that time, but they proved to be cacoxenite and strengite, which had not been previously reported from Arkansas. Since that time, a total of eight iron phosphates plus associated minerals have been found at several localities in the area.

### Location and Geologic Setting

The principal localities for the iron phosphates are shown in Figure 1, on which the pre-Stanley rocks are shaded and depict the older, central part of the Ouachita Mountains. The exact locations are listed in the Appendix. They are all restricted to the Arkansas Novaculite which, due to its resistance to erosion, forms some of the most rugged topography in the Ouachitas.

Most of the localities are near the upper part of the massive lower division of the novaculite, where the phosphate minerals occur within fracture zones associated with goethite and various manganese oxides and hydroxides. Localities 1-6 occur within a strike length of 20 miles in about the same part of the novaculite sequence, although traces of phosphate mineralization also occur at other places along this trend and on either side of it. Locality 1 is where a coating of turquoise on the novaculite prompted the drilling of two churn drill holes for copper (circa 1954). Localities 2 and 3 are the north and south pits dug by Mr. Isom Avants. Localities 4 and 5 are both on Buckeye Mountain, no. 4 being an old prospect pit in turquoise-coated novaculite, and no. 5 a breccia filling of

iron phosphates encountered in the construction of an access road. Locality 6 is the old Coon Creek manganese mine, and locality 7 is Fodderstack Mountain, about 5 miles southwest of Norman. Additional search will no doubt lead to the discovery of more localities. The best places to search for the mineralization is in the many old prospect pits that have been dug, principally for iron and manganese. Other likely spots are in novaculite outcrops where the formation has been fractured or brecciated and the openings filled with goethite.

The no. 8 locality is at Wilson Springs in novaculite adjacent to the igneous intrusive in the Union Carbide vanadium pit. The mineralization here is quite different in appearance from that at the localities farther west and is the only place where aluminum phosphates have been found to occur with the iron phosphates.

### Previous Work

The earliest reference to iron phosphates in the area are by Miser (1917) who noted the occurrence of sparse "dufrenite" at some of the manganese prospects that he described. Frondel (1949) in his discussion of dufrenite noted that several different minerals had been included under this name, and he described and named two new minerals, rockbridgeite and laubmannite, both of which were reported to occur in Polk County, Arkansas. The description of laubmannite was based upon the Arkansas material, which was the only locality known at the time. Fisher (1966) reported X-ray diffraction and optical data for cacoxenite from Polk County, Arkansas. Moore (1970) published a description of the crystal structure of several basic iron phosphates and included X-ray powder patterns which revised those given by Frondel. The minerals reported by Moore from the area were laubmannite, rockbridgeite, dufrenite, lipscombite, beraunite, and mineral "A".<sup>2</sup>

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Manuscript received, June 1, 1973

<sup>2</sup> Editor's footnote--Mineral "A" has been described and named KIDWELLITE in honor of Dr. Albert L. Kidwell by Dr. Paul Brian Moore, University of Chicago, and the name has been approved by the International Commission on New Mineral Names. A more complete description of the new mineral by Dr. Moore is in press in the Mineralogical Magazine.

**LOCATION OF IRON PHOSPHATE OCCURRENCES  
OUACHITA MOUNTAINS, ARKANSAS**

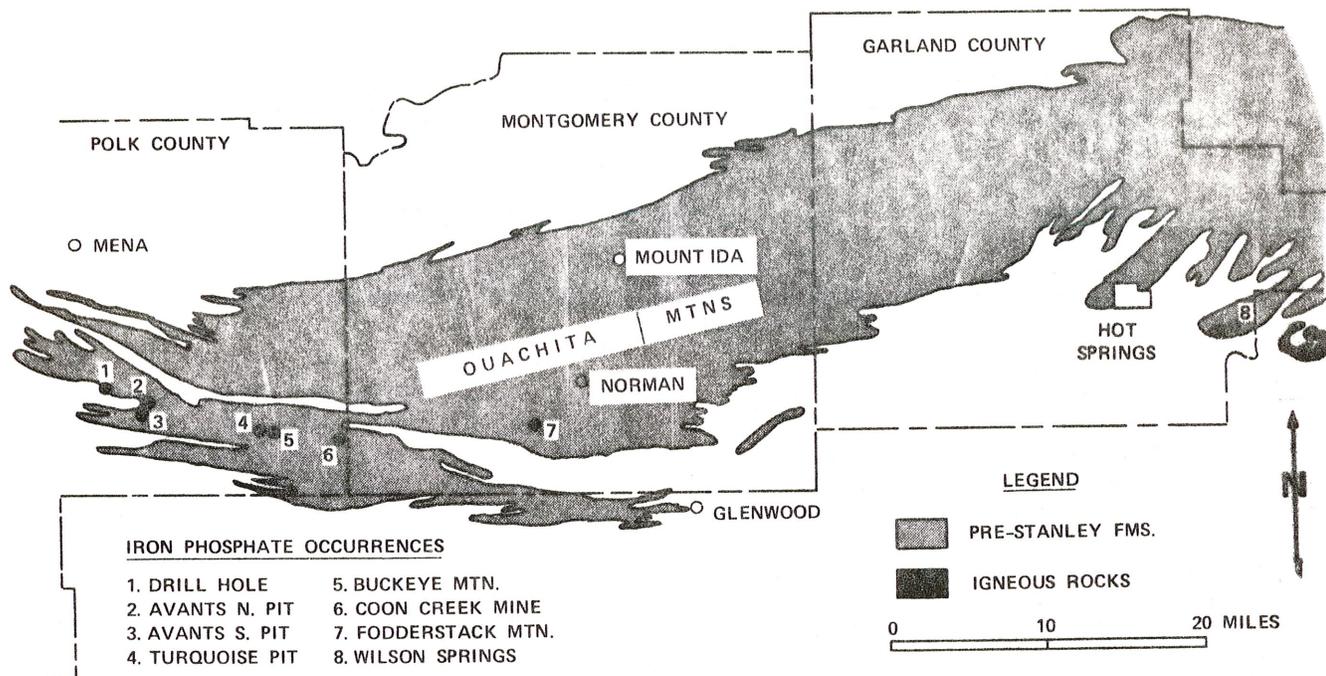


Figure 1. -- Location of iron phosphate occurrences Ouachita Mountains, Arkansas.

**DESCRIPTION OF INDIVIDUAL MINERALS**

The minerals will be described in their paragenetic sequence insofar as possible. However, this is only approximate, because sometimes it is impossible to tell which was first, and frequently there is cyclical deposition which is repeated several times. The earliest mineral to form was goethite, then the ferrous-ferric phosphates, followed by predominantly ferric phosphates in the later stages. No ferrous phosphates have been found. Likewise, no manganese phosphates have been found, nor have any been reported from this type of deposit.

Goethite was deposited prior to any of the phosphate minerals in the form of infiltrations of the novaculite, fracture fillings, and botryoidal or stalactitic growths in some of the larger openings. In the latter occurrences it has a definite fibrous texture. The present outer surface is darker in color than the inside and is frequently iridescent. In thin section, the centers of the stalactitic masses are frequently seen to be almost isotropic but with an outer layer of well-crystallized birefringent material. In addition to the earlier goethite, it also occurs recurrently as thin bands interlayered with the iron phosphates and as tiny spheres deposited on the other minerals.

Some of the manganese oxides and hydroxides are interlayered with the goethite, but no effort was made to work out their mineralogy. The manganese minerals may

also be late and form thin coatings on some of the iron phosphate minerals.

**Ferrous-Ferric Phosphate Minerals**

Several of the ferrous-ferric phosphates are very similar in their appearance, consisting of green, radiating fibers making up botryoidal aggregates. The botryoids invariably show banding, some of which is due to slight changes in the composition of a single mineral, and some due to the presence of different mineral phases. In many instances, the individual fibers change in mineral composition along their length. The most reliable means of specific identification is the X-ray diffraction pattern, and generally the position of the first X-ray peak in the 8 to 15 angstrom region is diagnostic. The next best criteria for identification are the color, pleochroism, and general appearance of the fibers.

Another rather disappointing property of these minerals is the relatively great ease with which they are altered by weathering. As a result, their color changes from green to various shades of brown, and the minerals end up mineralogically as goethite.

A resinous, reddish brown material is present at most of the localities and typically occurs at the base of some of the iron phosphate growths. The writer was not able to obtain a powder pattern with the X-ray spectrometer, but Moore (1970) reported that "the powder pattern of

some gels resemble goethite, and with broadened lines". Qualitative elemental analyses of several samples by X-ray fluorescence with the stereoscan microscope showed both iron and phosphorous to be present in roughly equal amounts. The material probably represents an iron phosphate gel that has never crystallized.

**Laubmannite:**  $\text{Fe}_3^{\text{II}}\text{Fe}_6^{\text{III}}(\text{PO}_4)_4(\text{OH})_{12}$  typically occurs as olive green to greenish brown fibers in radiating masses which also show color banding. The fibers tend to be coarser and more definitely bladed than do the other phosphates. They are highly pleochroic from yellow or orange-yellow to bluish-green. The banding appears to be due to (1) slight changes in degree of oxidation which may reflect differences in the original ferrous-ferric ratio and (2) the presence of thin layers of goethite, dufrenite and/or mineral "A", especially near the outer surface of the botryoids. Typical specimens of laubmannite are shown in Figure 2, A and B. The fibrous and banded nature of the mineral is shown in Figure 3, A, which is a photomicrograph of a thin section of laubmannite. Although laubmannite occurs throughout the area, the best localities at present are Buckeye Mountain and the Coon Creek mine.

**Rockbridgeite:**  $\text{Fe}_4^{\text{II}}\text{Fe}_3^{\text{III}}(\text{PO}_4)_3(\text{OH})_5$  is distinguished principally by the color, which ranges from greenish-black when fresh to brownish when oxidized. The fresh material is distinctly darker than the other minerals with which it might be confused. It generally occurs in either radiating fibrous masses or in fibrous veins within the iron-impregnated novaculite. Individual crystals sometimes occur lining cavities between intersecting groups of botryoids. Optically, the fibers are highly pleochroic from pale yellow-brown to bluish-green. Rockbridgeite and beraunite frequently occur together, and the composition may change from one to the other along a single fiber, as at the Avants south pit. At the Fodderstack Mountain locality, rockbridgeite shows several cycles of deposition with lipscombite, mineral "A", and goethite (Figure 3, B).

**Lipscombite:**  $\text{Fe}_2^{\text{II}}\text{Fe}_2^{\text{III}}(\text{PO}_4)_2(\text{OH})_2$  was first recognized by Moore (1970) from the Fodderstack Mountain locality. It is the only bluish-green mineral in the suite, and forms granular layers in association with layers of rockbridgeite and mineral "A". So far it has not been found at any of the other localities.

**Beraunite:**  $\text{Fe}_5^{\text{II}}\text{Fe}_5^{\text{III}}(\text{PO}_4)_4(\text{OH})_5 \cdot 6\text{H}_2\text{O}$  is one iron phosphate that is easily recognized on the basis of its color. In the Arkansas specimens it is invariably a characteristic orange-red to hyacinth-red. It occurs in granular to prismatic aggregates but not usually in fibrous masses except locally where it replaces rockbridgeite (Figure 4, A). Groups of tabular crystals up to a centimeter in length are sometimes found in cavities. Optically, the beraunite is very distinctive in its pleochroism from pale yellow to reddish brown. Beraunite is found at most of the localities and occurs with both rockbridgeite and laubmannite.

**Dufrenite:**  $(\text{Fe,Ca})_3\text{Fe}_{10}^{\text{III}}(\text{PO}_4)_8(\text{OH})_{12} \cdot 4\text{H}_2\text{O}$  was first noted by Moore (1970) from the Buckeye Mountain locality where it occurs as brilliant green microscopic crystals that terminate fibers of laubmannite. The botryoidal aggregates of laubmannite from the Coon Creek mine also commonly have a bright emerald-green band near the ends of the fibers that is probably dufrenite. This is a tentative identification based on color, because the writer was not able to isolate enough pure material for an X-ray pattern with the X-ray spectrometer. This appears to be one of the rarer minerals of these localities.

**Mineral "A"** (see editor's footnote<sup>2</sup>) is a ferrous-ferric phosphate that was noted by Moore (1970) who reported its X-ray diffraction pattern and general appearance. It is presently being described and named by him, and a complete description will appear in print in the near future. Binocular microscope and thin section study and numerous X-ray diffraction analyses by the writer have shown that it is one of the more common iron phosphates in the area. In addition to the X-ray pattern, mineral "A" is characterized by its color, which may be various shades of light green and greenish yellow, and the relatively weak pleochroism from light yellow to light yellowish green. Interference colors are typically an anomalous Berlin blue.

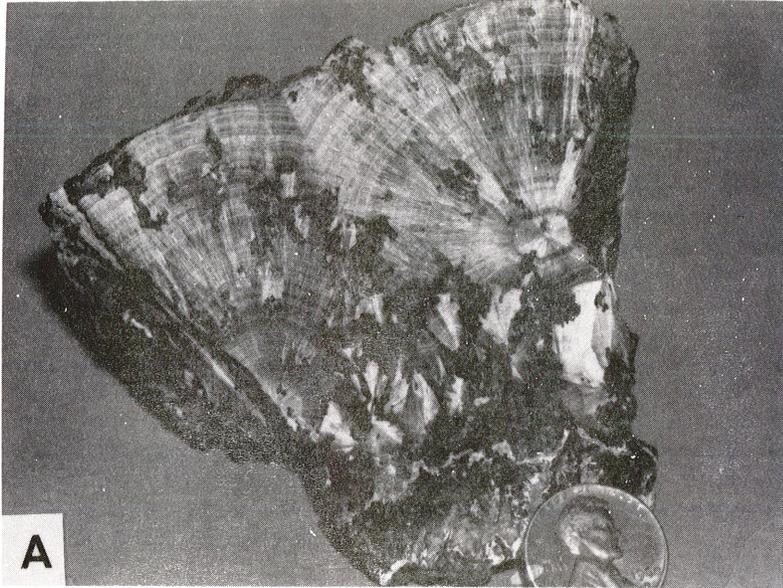
Three types of occurrences are typical for mineral "A": (1) interlayered with laubmannite and rockbridgeite in botryoids and particularly forming the outer layer, (2) forming thin, fibrous botryoidal masses of its own, and (3) as single crystals, sheaves of crystals, or individual spheres up to 1-2 mm in diameter on goethite. The interlayering with laubmannite is shown in the photomicrograph of Figure 4, B, taken of a thin section in which there are numerous alternating bands. The nature of the individual crystals and crystallized aggregates is shown in four photomicrographs taken with a scanning electron microscope. Figure 5, A, illustrates a typical botryoidal mass with radiating crystals, which is characteristic of many phosphate minerals. Crystal terminations are present but cannot be seen very clearly at this magnification. Figure 5, B, shows a 2000X enlargement of one of the fibrous crystal terminations with smaller crystals at right angles. This feature was noticed in several instances and gives the appearance of true twinning. Figure 6, A, depicts another variation in which the crystals are very fibrous and with very indefinite terminations, even at much greater magnifications than that shown here. A fourth variation (Figure 6, B) consists of bundles of crystals on a botryoidal surface of goethite.

Some of the material probably represents an alteration in place of laubmannite or rockbridgeite, but much of it is definitely a result of later solution and redeposition.

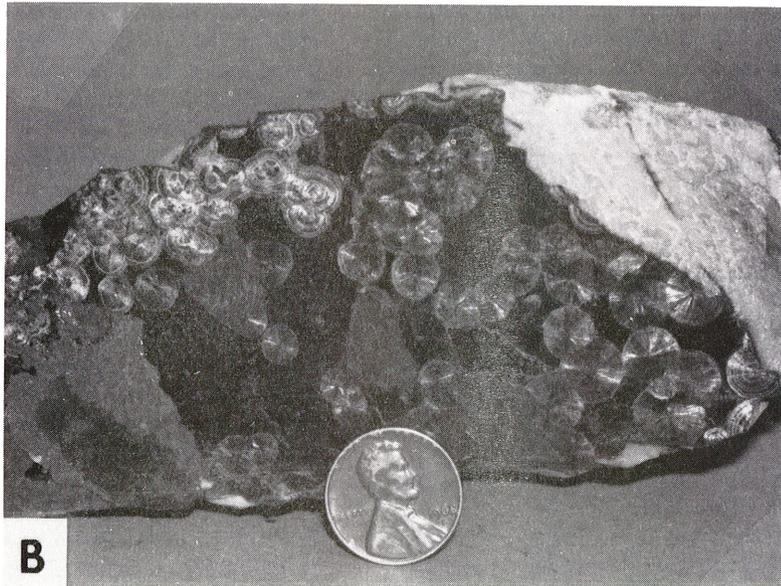
#### Ferric Phosphate Minerals

The next group of minerals to be considered are the ferric phosphates, strengite and cacoxenite. These species are much more stable than the ferrous-ferric phosphates and invariably occur later in the paragenetic sequence.

**FIGURE 2**

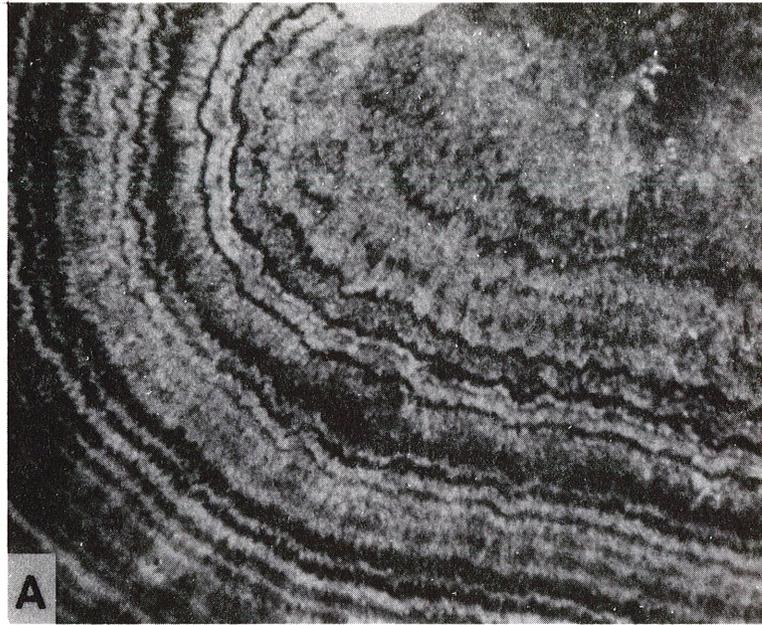


**A.** Laubmannite interlayered with dufrenite and mineral "A", Buckeye Mountain, actual size.



**B.** Laubmannite spherulites with thin layers of mineral "A", Coon Creek mine, actual size.

FIGURE 3

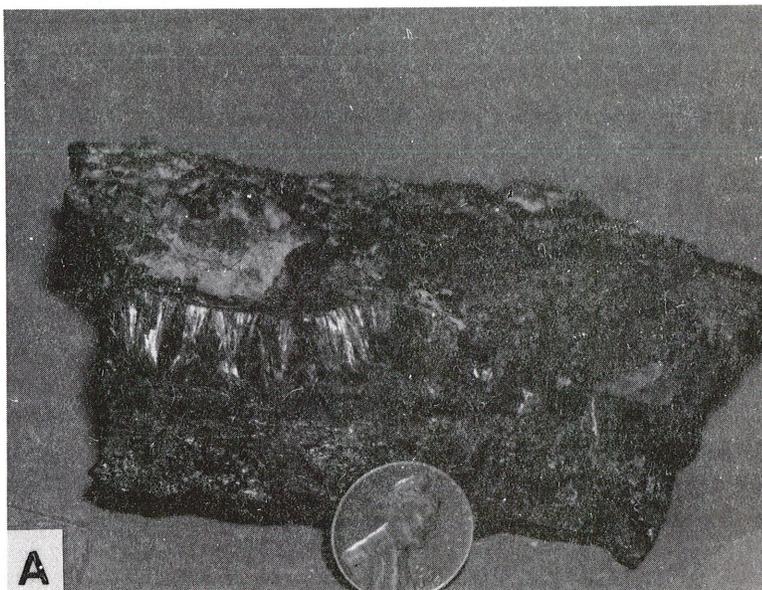


A. Interlayered laubmannite and goethite shown in thin section, Buckeye Mountain, 35X, ordinary light.

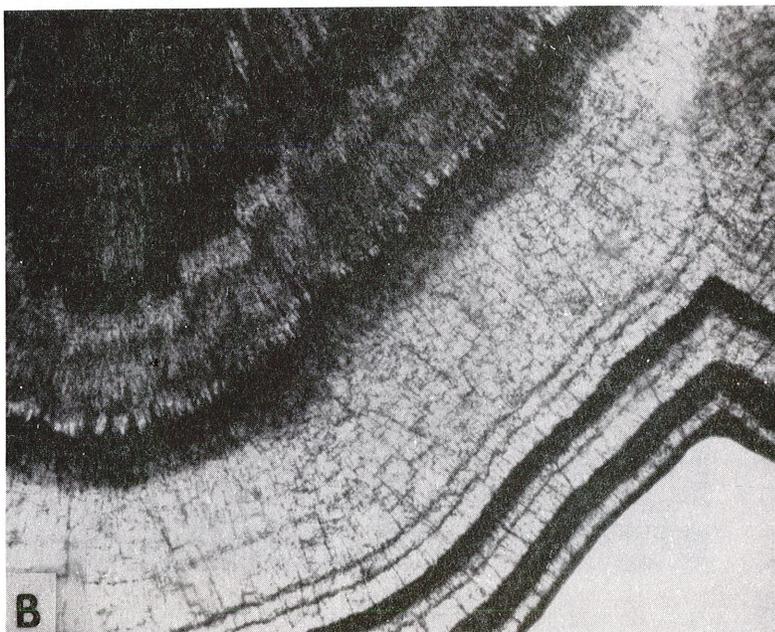


B. Interlayered rockbridgeite, lipscombite and mineral "A" (light) showing cyclical deposition, Fodderstack Mountain, actual size.

**FIGURE 4**



**A. Fibrous layer of beraunite replacing rockbridgeite, Avants south pit.**

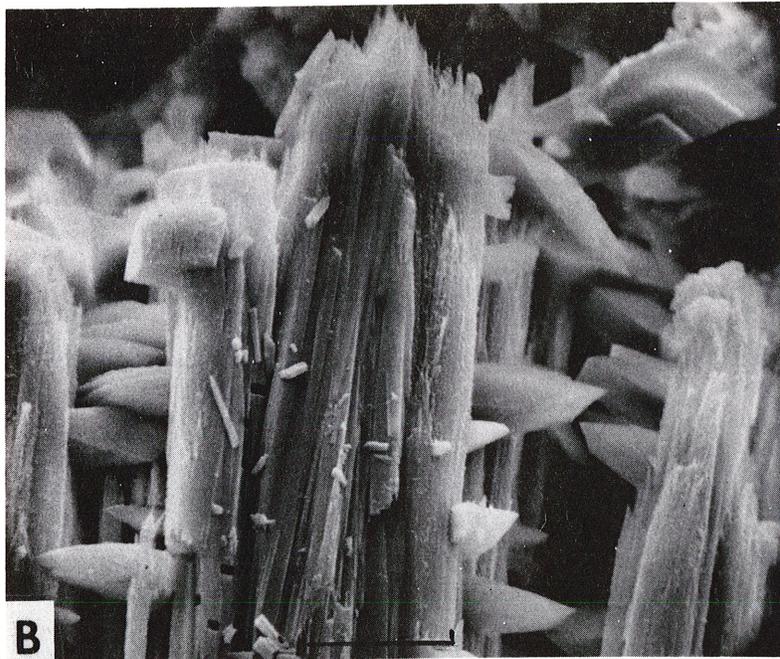


**B. Bands of mineral "A" (light gray) interlayered with laubmannite (dark gray) and goethite (black), Coon Creek mine, 35X, ordinary light.**

FIGURE 5

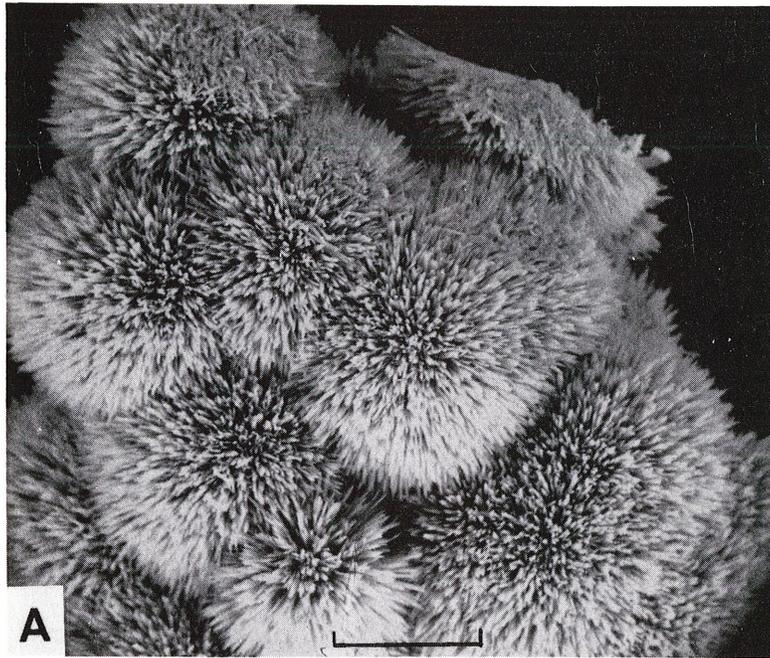


**A.** Mineral "A", botryoidal mass of radiating crystals, Coon Creek mine, 50X, scale = 200  $\mu$ , scanning electron microscope.

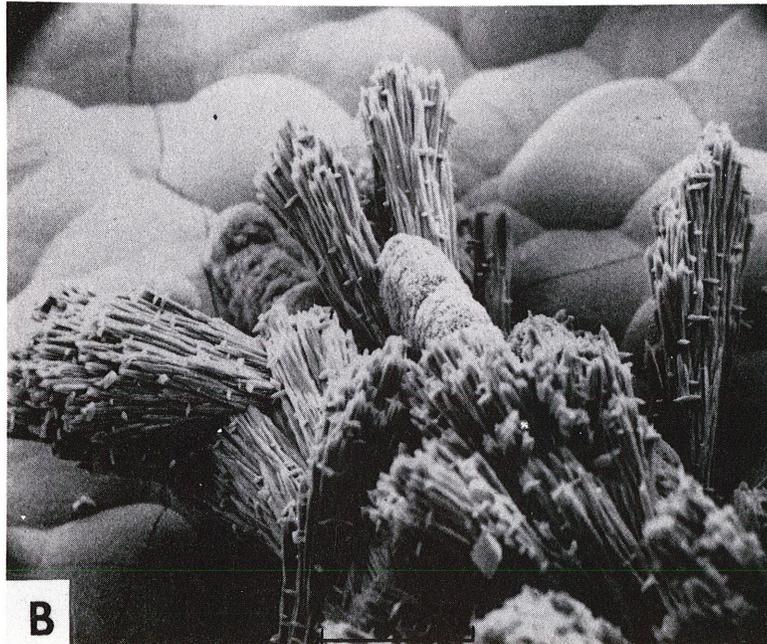


**B.** Same as above but with 2000X magnification, scale = 10  $\mu$ . Note small crystals oriented normal to main fibers.

FIGURE 6



A. Mineral "A" showing fibrous crystals with indefinite terminations, Coon Creek mine, 200X, scale = 200  $\mu$ , scanning electron microscope.



B. Mineral "A" showing sheaves of crystal fibers with smaller crystals at right angles, botryoidal surface is goethite, Coon Creek mine, 200X, scale = 100  $\mu$ , scanning electron microscope.

**Strengite:**  $\text{FePO}_4 \cdot 2\text{H}_2\text{O}$  typically occurs as globular masses up to one-half cm in diameter of radiating pink, gray, or violet terminated crystals, or it may spread out as flat, rosette-like masses along fracture zones. Groups of well-formed bladed crystals also occur in many of the vugs. Still another type of occurrence is in the form of smooth, pearly, light gray spheres up to 1-2 mm in diameter and with concentric layers but no individual terminations. Figure 7, A, shows a small vug lined with radiating groups of rose-colored crystals. Strengite is commonly associated with cacoxenite, and the two minerals are particularly abundant at the Avants north pit and at the Coon Creek mine. Strengite also occurs at the Wilson Springs locality in the form of pink to tan, micro-spherical aggregates on a layer of cacoxenite.

**Cacoxenite:**  $\text{Fe}_9(\text{PO}_4)_4(\text{OH})_{15} \cdot 18\text{H}_2\text{O}$  is the most easily recognized and most widespread iron phosphate of the entire group. It occurs in the form of yellow, fibrous crystals which may be up to 1-2 cm long or may form small botryoids composed of radiating fibers. The cacoxenite is sometimes found growing directly on patches of the amorphous iron phosphate gel. Figure 7, B, shows typical fibrous crystals growing into a vug in goethite. Figure 8, A and B, are photomicrographs taken with the scanning electron microscope and show the nature of the cacoxenite fibers very graphically. Terminations, where they can be seen, are somewhat rounded and indistinct. As mentioned earlier, cacoxenite and strengite typically occur together, although cacoxenite is somewhat more widely distributed away from the localities shown on the map.

#### Associated Minerals

A secondary mineral commonly associated with the iron phosphates is **turquoise**  $\text{CuAl}_6(\text{PO}_4)_4(\text{OH})_8 \cdot 4\text{H}_2\text{O}$  and/or the closely related material that has been called **planerite**  $\text{CuAl}_6(\text{PO}_4)_4(\text{OH})_8 \cdot 3.7\text{H}_2\text{O}$ . The composition of the two is virtually identical, but there are some differences in their X-ray diffraction pattern and optical properties. The two minerals are considered to be different species by Fischer (1958) and by Cech et al., (1961). However, Fleischer (1962) noted that the evidence does not seem convincing and that further study is needed. Most of the material on which the writer has run X-ray powder patterns has the properties reported for planerite, but there appear to be gradations between the two minerals. The minerals occur typically in the form of botryoidal and stalactitic coatings along fractures in the novaculite. The color ranges through various shades of brown, green, and bluish-green and have a waxy luster. A single specimen was found at the Coon Creek mine in which microcrystals of definite turquoise are coating goethite as shown in Figure 9, A and B. These are photomicrographs taken with the scanning electron microscope and show the typical crystal form of turquoise. The X-ray fluorescence spectra of the crystals shows the presence of aluminum, copper, iron, and phosphorous, which confirmed the chemical composition of turquoise.

The turquoise and planerite (if it is a separate species) have probably formed by near-surface alteration of the iron phosphate deposits, because (1) they occur in or adjacent to all of the iron phosphate localities, and (2) in all of the turquoise localities there are relicts of the typical radiating fibrous structure of the iron phosphates.

**Wavellite:**  $\text{Al}_3(\text{OH})_3(\text{PO}_4)_2 \cdot 5\text{H}_2\text{O}$  and **variscite**  $\text{Al}(\text{PO}_4) \cdot 2\text{H}_2\text{O}$  have not been found with the iron phosphates except at Wilson Springs where they occur as botryoidal layers and microcrystals along fractures. Likewise, none of the iron phosphates have been reported to occur at any of the well-known wavellite and variscite localities, such as the ones at Avant (Buckville PO), Arkansas.

**Hyalite opal:**  $\text{SiO}_2 \cdot n\text{H}_2\text{O}$  is a minor but widely distributed component of the iron phosphate deposits. It occurs as typical thin, colorless coatings on goethite and the phosphate minerals. None of the samples tested have shown any fluorescence under the short wavelength lamp. In one sample, plant roots were found to be coated with opal, indicating that at least part of it is very recent.

#### ORIGIN OF THE DEPOSITS

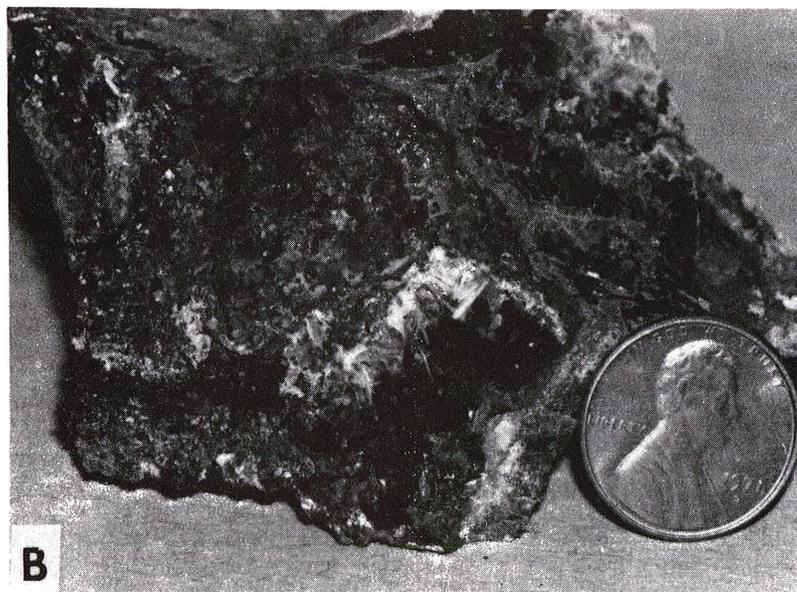
No geochemical work has been done on the novaculite or adjoining formations to determine the source of the phosphate. However, the virtual restriction of the iron phosphate deposits to the lower and middle parts of the novaculite suggests that the source is in or slightly above this part of the section. One possible source is the conodonts and linguloid brachiopods that are known to occur within the middle division of the novaculite (Miser and Purdue, 1929, p. 58). Small amounts of phosphorous are present in all of the manganese samples from the novaculite that have been analyzed.

The solutions that carried the phosphate were probably the same ones that deposited the iron and manganese in fracture zones of the novaculite. There is no evidence to suggest that they were other than meteoric water that deposited its mineral load in a succession of colloidal layers that later crystallized to form the ferrous-ferric phosphates. There is abundant evidence from thin-section examination of periodic interruptions in the depositional process and of cyclical deposition due to recurring changes in the chemical environment. The ferric phosphates were formed later and nearer to the land surface by oxygenated waters that leached part of the earlier minerals and redeposited the iron and phosphorous as secondary minerals. Late in the alteration process small amounts of copper were leached from the novaculite and combined with iron, phosphorous, and aluminum to form turquoise as one of the last minerals. Thus, the suite of iron phosphate minerals described in this paper represents a classic example of the adjustment of a mineral assemblage to a changing chemical environment.

FIGURE 7



A. Crystallized rosettes of strengite along fracture in novaculite, Avants north pit.

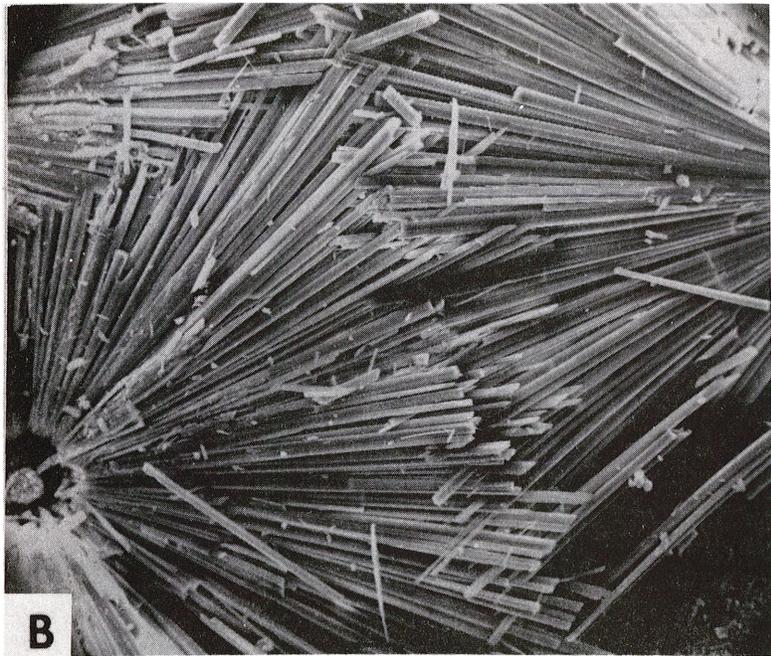


B. Fibrous crystals of cacoxenite in vug lined with goethite, Avants north pit.

FIGURE 8

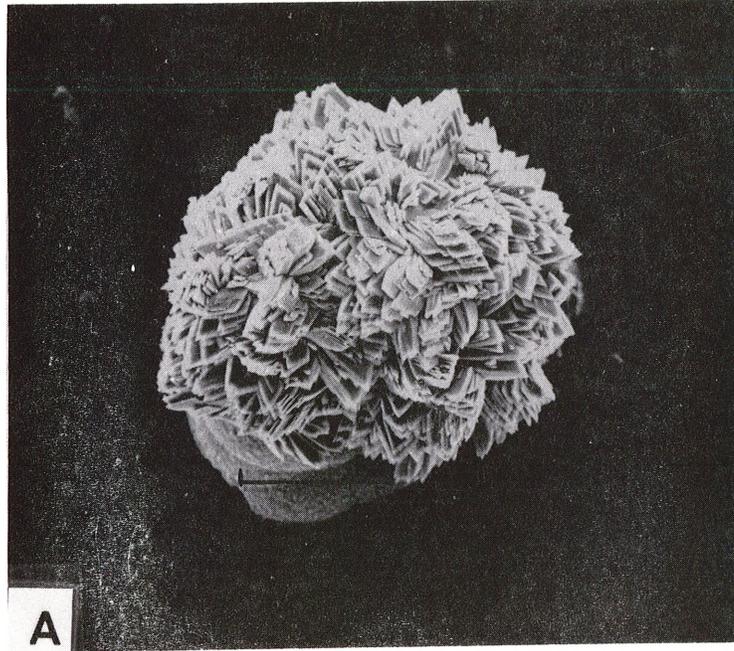


A. Fibrous cacoxenite crystals in goethite-lined vug, 50X, scale = 200  $\mu$ , scanning electron microscope

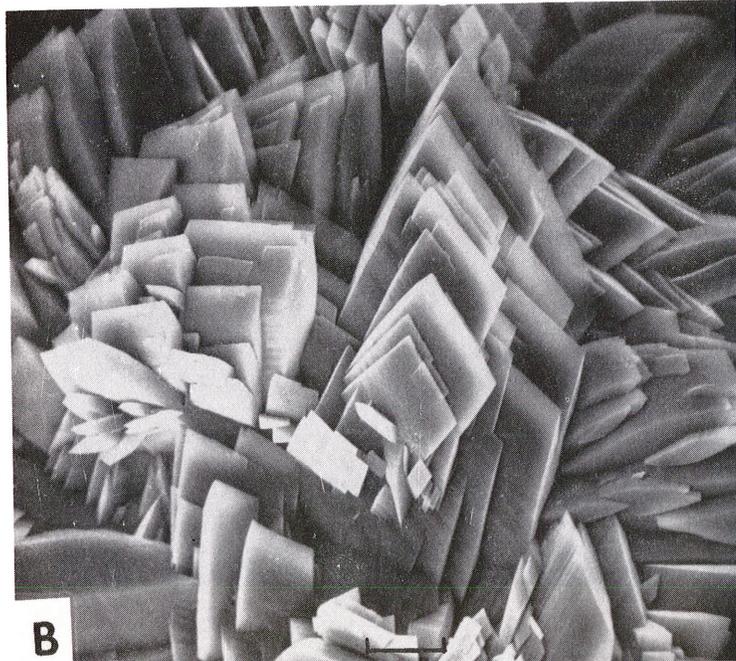


B. Intergrown cacoxenite crystals from different centers of crystallization, 200X, scale = 100  $\mu$ , scanning electron microscope.

FIGURE 9



A. Cluster of turquoise crystals on goethite base, Coon Creek mine, 200X, scale = 100  $\mu$ , scanning electron microscope.



B. Same as above but with 2000X magnification, scale = 10  $\mu$ .

## ACKNOWLEDGEMENTS

The writer wishes to thank the management of Exxon Production Research Company for permission to publish this paper and for the use of company equipment utilized in identification and description of the minerals. Special

acknowledgements also go to Mr. Henry deLinde, Hot Springs, for his assistance and encouragement and to Mr. Isom Avants, Mena, for his cooperation in digging prospect pits at exactly the right places.

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## APPENDIX

### Iron Phosphate Locations

1. Center SW $\frac{1}{4}$  SW $\frac{1}{4}$  SE $\frac{1}{4}$  sec. 34, T. 3 S., R. 30 W. An abandoned road leads north from the Holly Branch road and then turns east to the spots where two shallow churn drill holes were put down (circa 1954) on the south side of the WNW-ESE trending novaculite ridge. Fracture coatings of turquoise are common in the novaculite, and cacoxenite has also been found in fractures.
2. SE $\frac{1}{4}$  SE $\frac{1}{4}$  NW $\frac{1}{4}$  sec. 1, T. 4 S., R. 30 W. A pit was dug by Mr. Isom Avants in the bed of the south-flowing ravine where a resistant E-W striking bed of novaculite containing iron phosphates crosses the stream. This is one of the best localities for collecting cacoxenite and strengite. Other minerals that have been found here are beraunite, laubmannite, mineral "A", and hyalite opal.
3. SE $\frac{1}{4}$  SE $\frac{1}{4}$  SE $\frac{1}{4}$  sec. 2, T. 4 S., R. 30 W. A trench was dug by Mr. Isom Avants in a north-sloping spur to West Hanna Mountain. The shaly middle division of the novaculite occurs here and is highly impregnated with goethite. This is the best locality for rockbridgeite, fibrous layers of beraunite replacing rockbridgeite, and the reddish brown iron phosphate gel. Other minerals found here are cacoxenite, strengite and hyalite opal.
4. NE corner NE $\frac{1}{4}$  NW $\frac{1}{4}$  sec. 18, T. 4 S., R. 28 W. This is an old prospect pit dug into turquoise-coated, fractured novaculite. From the spot where the road crosses Buckeye Mountain take the abandoned mine access road to the west until it turns to the south. The pit is in one of the steep ravines to the north, about 100 feet above the road. This is the best locality for turquoise-planerite, but all of the iron phosphate minerals have been completely altered.
5. NW $\frac{1}{4}$  NW $\frac{1}{4}$  sec. 18, T. 4 S., R. 28 W. The original discovery of laubmannite was made along the mine access road just west of the main road, but all of the obvious mineralization has been carried off. However, the same type of material occurs about 100 yards to the east on the narrow spur that leads to Blaylock Mountain where the novaculite contains goethite and iron phosphates in brecciated zones. In addition to laubmannite, there is abundant strengite, beraunite and hyalite opal.
6. NW $\frac{1}{4}$  SE $\frac{1}{4}$  NW $\frac{1}{4}$  sec. 13, T. 4 S., R. 28 W. This is the abandoned Coon Creek mine that produced a small amount of manganese during the 1950's. The only access by road is from the east by the primitive road that crosses Little Missouri River in sec. 21, T. 4 S., R. 27 W. and follows Long Creek valley to the west. The road should not be attempted without a truck or other field vehicle. The mineralization occurs in fractured novaculite within and on both sides of a steep ravine that cuts across strike. All of the phosphate minerals with the exception of lipscombite have been found here.
7. SW $\frac{1}{4}$  SW $\frac{1}{4}$  sec. 7, T. 4 S., R. 25 W. and SE $\frac{1}{4}$  SE $\frac{1}{4}$  sec. 12, T. 4 S., R. 26 W. There are several small prospect pits and other unprospected areas of iron phosphates with goethite in the novaculite on the south side of the mountain. This is a densely forested area, and it is impossible to give specific instructions. The original samples in which mineral "A" was found were from a small pit near the crest of the Fodderstack range, and this is the only locality where lipscombite has been found. Most of the other iron phosphates occur at one place or another in the area.

# ANALCITE AS A TEMPERATURE INDICATOR OF ARKANSAS

## OUACHITA DEFORMATION

By Kern C. Jackson<sup>1</sup>

A preliminary study of the zeolite minerals in the Carboniferous rocks of the Ouachita Mountain area of Arkansas was undertaken in the summer of 1966 with the hope of developing a thermal zonation of that area through the zeolite—and possibly lowest greenschist—facies of metamorphism. A number of zeolite minerals are known to develop during diagenesis of sediments. However, a very restricted assemblage of zeolite minerals is stable under metamorphic conditions when temperatures reach or exceed 300° C and none are stable above 400° C. The zeolite minerals found in this study were all those of diagenetic origin, indicating that the samples studied were never exposed to sustained temperatures above 300° C.

Samples were collected from the Stanley and Jackfork Formations in three areas. Four samples were collected from an area north and northwest of Little Rock, Arkansas toward the Y City thrust zone. A second suite of seven specimens was collected along Arkansas Highway #7 to the north and south of Hot Springs, Arkansas scattered over a north-south distance of forty miles. The third set of six specimens was collected from outcrops along and near U. S. # 71 along the Arkansas-Oklahoma border from DeQueen to Mena, Arkansas, a north-south distance of thirty-six miles. The specimens studied were selected in the field as graywacke or subgraywacke sandstones. Quartz veins are abundant in many of the outcrops and samples were selected as far from veins as possible.

Thin section study of the seventeen samples show fourteen to be graywacke with abundant angular to rounded quartz as the dominant sand-sized material. Potassium feldspar, plagioclase in the oligoclase to andesine range of composition, mica flakes, and shale and phyllite rock fragments are common in the sand-sized material especially in specimens from the eastern localities. An argillaceous matrix makes up 20 to 60% of the material in most specimens. The other three specimens proved to be sandy carbonate rocks and were apparently either lenses or concretions in the shales. The carbonate was thoroughly recrystallized and extensively replaced the margins of the quartz sand grains. No zeolite minerals were ever recognized in the thin sections.

From each sample approximately two pounds of rock was prepared. All weathered surfaces were cobbled off and thin quartz veins were removed before crushing. Repeated crushing through steel rolls and screening on a Ro-Tap removed grains as they were liberated. About twenty (20)

grams of the -140+200 mesh size were separated on bromoform adjusted to a specific gravity of 2.4 with acetone. The light concentrate was studied in detail.

Analcite was found in the light concentrate of every sample ranging in amount from three grains to two dozen grains per sample. Two distinct habits are present. The most common form consists of bizarre shaped grains which appear to be concretionary aggregates of crystals. A sufficient amount of material for X-ray determination was never recovered. The identification of analcite is based on optical and microchemical techniques. The grains are pale pink, completely isotropic, and have an index of refraction between 1.484 and 1.486. A grain dissolved slowly under the microscope in a drop of 50% HCl and left no residue. Evaporation of the drop yielded water soluble, hopper-shaped cubic crystals, probably NaCl; and anisotropic prismatic crystals with parallel extinction which were delequiescent and water soluble. These were probably AlCl<sub>3</sub>. The total chemical behavior is that of analcite. Complete aggregates and fragments of such aggregates are consistently present in each sample.

The second habit of analcite consists of rounded grains of somewhat darker pink color and apparently higher negative relief when mounted in canada balsam. The grains appear to be extensively cracked, but under high magnification can be seen to have a rough pitted surface. The pits appear to be small rounded depressions on the surface and may be impressions of adjacent clastic grains on analcite grown in voids. The chemical behavior of these grains is identical to the bizarre shaped grains. They are less common occurring in only six samples.

Thomsonite is the only other zeolite mineral present and occurs in only two specimens, one from a graywacke immediately underlying the Hatton Tuff and the other from a tuffaceous appearing bed in Hot Springs city limits. The mineral appears as euhedral, subparallel, bladed crystal clusters. Neither index of refraction nor microchemical tests were made as there was only one grain in each specimen. Optic sign, 2V, and optic orientation were determined and fit thomsonite.

Coal is a minor to the major constituent of the light fraction of all specimens. In the crushed samples most coal appears as vitreous lustered, opaque chips with well developed conchoidal fracture. In the thin sections coal is usually present as irregular granules and flakes, but a cellular woody texture is preserved in one thin section. At several localities on the outcrop similar cellular texture can be seen on some bedding surfaces.

Analcite is a common mineral in some sediments. Three modes of origin have been recognized by Coombs

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and Whetten (1967). "Group A are silica-rich analcimes formed from siliceous volcanic glass by inferred reaction with saline waters. Group B are analcimes occurring in a burial metamorphic environment. Group C are silica-poor analcimes inferred to be formed by direct precipitation or by reaction of highly alkaline water with sediments." Analcites found in the present study are interpreted as belonging to Group B inasmuch as the samples are from a highly deformed thick sequence of unmetamorphosed sediments.

Coombs, Ellis, Fyfe, and Taylor (1959) have investigated the reaction:



At 500 bars water pressure analcite yields albite at 275° C. With increasing pressure the reaction temperature rises to a maximum of 295° C at 5,000 bars. At higher pressure the reaction temperature decreases slightly. Campbell and Fyfe (1965) investigated the analcite-albite equilibrium and found the equilibrium temperature at  $190 \pm 10^\circ$  C at 12 bars. Campbell and Fyfe were working with the low temperature - ordered - form of albite. They point out that if analcite were to form from disordered albite in solid solution in either plagioclase or alkali feldspar, the equilibrium temperature for the reaction would be raised.

This suggests that the data of Coombs et al., with equilibrium temperatures in the range of 275-295° C is in error by no more than a few tens of degrees. The error arising from the difficulty of obtaining equilibrium in reactions involving the zeolites. Turner (1968, p. 157) plots this equilibrium curve as an extrapolated curve based on the data of Campbell and Fyfe with a markedly decreasing equilibrium temperature at increased pressure. His curve passes through points 140° C at 3 k bar and 100° C at 3.6 k bar.

Analcite in the Stanley-Jackfork is always associated with abundant quartz and detrital feldspars of igneous or metamorphic origin. Careful examination of the thin sections shows no authigenic overgrowth of albite on detrital feldspar. Inasmuch as the detrital feldspars are more or less altered, it is assumed that analcite formed by reaction of connate brines with feldspar during burial metamorphism or as a result of high fluid pressures during gravity slide deformation as described by Viele (1968). Thus the temperatures attained during the deformation of these units did not reach the zeolite facies metamorphism and were less than 300° C based on the data of Coombs et al., or less than 150° C based on the interpretation of Turner. This is compatible with the structural interpretation of Viele.

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PORE FLUIDS AS THE PRIMARY RHEOLOGIC CONTROL  
IN THE DEFORMATION OF THE EASTERN OUACHITAS

By John F. Karlo<sup>1</sup>

ABSTRACT

Observations of deformation styles seen in outcrop and interpretations of regional structural geology require that during deformation, the rocks of the Eastern Ouachita fold belt must have possessed a high degree of ductility. However, such ductile behavior is inconsistent with our understanding of the deformation mechanics of sub-greenschist metamorphic grade rocks. Thin section observations of intensely deformed rocks show evidence of syntectonic pore pressures and observed deformation modes are consistent with the hypothesis that water was the controlling rheologic factor during Ouachita deformation. Diagenetic dewatering of clays is proposed as the source of such fluids with the timing of the process related to the overall geosynclinal history of the orogen.

INTRODUCTION

The tectonics of the Eastern Ouachitas have been interpreted (Viele, 1966, 1973) as a polyphase deformational system of nappe formation caused by gravity loading followed by backfolding of the nappes and a period of regional arching. In conjunction with this interpretation, the rocks of the Eastern Ouachitas are divisible into two structural super-members: an infrastructure of Lower Paleozoic rocks whose "ordered and penetrative structures" result from deformation under confinement, and a supra-structure of Carboniferous rocks which passively blanketed the infrastructure and was deformed by infolding during nappe involution and backfolding (Viele, 1973).

Regional metamorphism has not been considered important to Ouachita deformation since the observed metamorphic grade does not exceed incipient greenschist facies (Flawn, 1961). However, it follows that since temperature and pressure conditions during deformation were very low, the ductility implied in the nappe hypothesis is extreme compared to other rocks of similar sub-greenschist metamorphic grade (Holland and Lambert, 1969). Nevertheless, the implied degree of ductility is consistent with the observed structural elements of the Eastern Ouachitas which include pervasive slaty cleavage, large scale isoclinal passive flow and flexural flow folds, and mullion and boudinage structures.

This apparent contradiction between expected and observed degrees of ductility raises the question that if temperature and pressure effects are excluded as the causes of macroscopic ductility, by what mechanisms did the rocks undergo ductile deformation and what features of the deformational environment were the controlling factors?

In order to gather information relating to this problem, the deformation mode was investigated in rocks sampled from varied lithologic types and small scale structural forms in a limited area where the overall stratigraphic and

structural relations were well known to the author. The study area (Fig. 1) occupies the southern half of the U. S. Geological Survey 7 1/2' Fourche SW quadrangle and the southeast corner of the Paron quadrangle within Saline and Pulaski Counties, Arkansas (Karlo, 1973). This area is situated along the northern flank of the Benton-Broken Bow Uplift and has exposed within it a series of digitate folds of the nappe-infrastructure and Carboniferous rocks of the suprastructure (for geologic map, see Viele, 1973).

RHEOLOGY

Deformation Mechanics

Macroscopic strain styles may provide excellent constraints upon the modes of rock deformation. However, since the recognition of such styles is highly interpretive and the macroscopic behavior is the summation of deformations on the level of component grains, the microscopic observations are more direct and much less subjective. Accordingly, this discussion will be limited to strain features as seen in standard thin sections.

The behaviors which can theoretically account for macroscopic ductility fall into three categories: (1) deformation by intergranular slip, (2) deformation by cataclastic intergranular slip, and (3) plastic deformation. Deformation by intergranular slip accommodates strain in a volume of rock solely by changes in shape of that volume produced by slip between the constituent grains. The grains themselves remain undeformed and the degree of ductility of the rock as a whole is directly related to the ease of intergranular slip as controlled by friction. As internal friction increases, the rock progressively loses ductility and eventually grades into cataclastic flow (Byerlee, 1968). Cataclasis occurs when the stress necessary to overcome friction is of sufficient magnitude that it exceeds the fracture strength of the grains. During cataclastic flow, a large component of the deformation is contributed by intergranular slip but during the process, grain fracturing and brecciation occurs. Ideally, there is no contribution by plastic flow (Borg, et al., 1960) and the behavior on the grain level is brittle in nature. With continued increase in internal friction, intergranular slip becomes impossible and the rock will show no macroscopic ductility until the conditions of the environment exceed the brittle-ductile transition for the constituent minerals. With the brittle-

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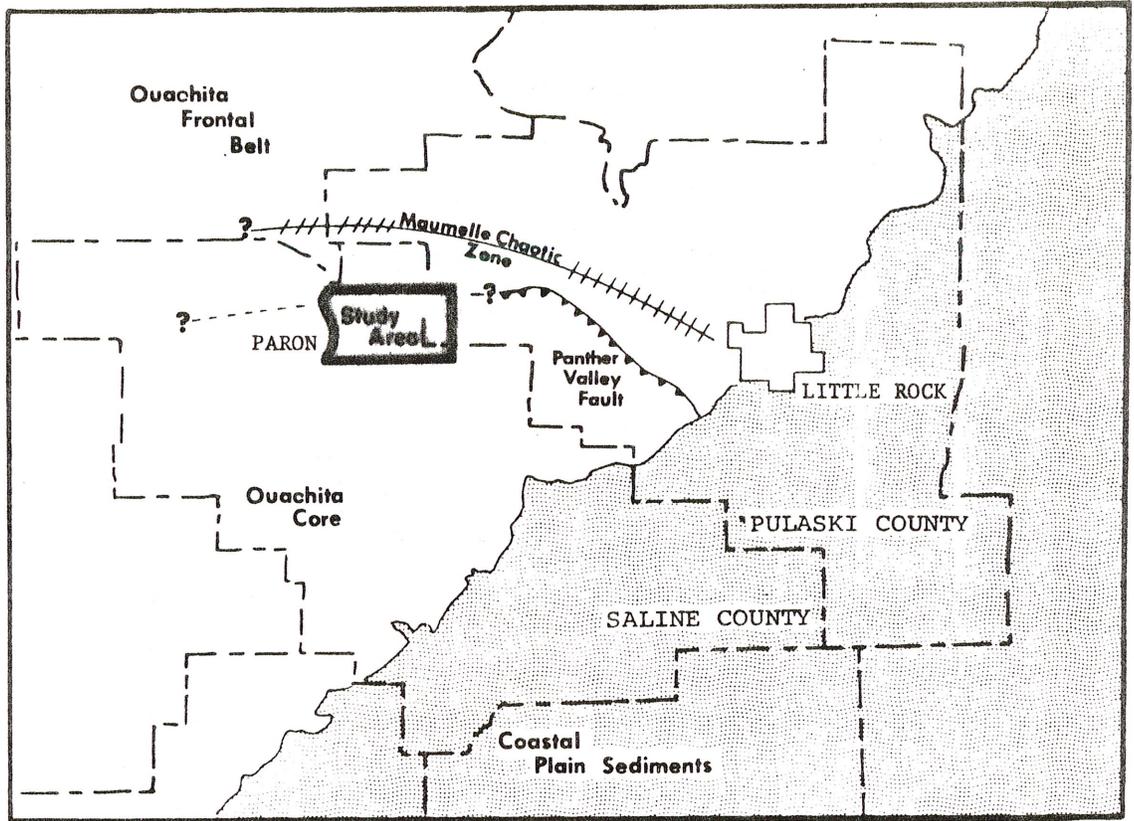


Figure 1. -- Location Maps of Study Area

ductile transition surpassed, rock deformation proceeds by plastic flow. Plastic flow in rocks occurs by distortions of grain crystal structure by such processes as twinning, dislocation movement, and recrystallization. By and large, plastic flow is the type of ductility normally associated with rock deformation, but it is expected to play a minor role in the deformation of sub-greenschist grade rocks (Holland and Lambert, 1969).

Although there are recognizable differences as functions of behavior, the macroscopic deformation features produced by the above mechanisms are essentially similar and any of these behaviors or combination thereof might have been active within the Eastern Ouachitas. Thus, a determination of what the dominant deformation mode was has a primary significance to understanding the environment of deformation.

### Observed Micro-strain Features

Rocks ranging from the lower Ordovician Mazarn to the Carboniferous Jackfork Formations are exposed within the Paron-Fourche SW area (for reviews of stratigraphy, see Cline et al., 1959; Goldstein, 1961). These formations are dominated within the study area by four lithologic groupings: limestones; cherts and siliceous shales; sandstones and shaley sandstones; and shales. Samples were collected of these lithologies only from outcrops with very well-developed macrostrain features, *e.g.* folds, boudins, bed brecciation, etc. The philosophy behind this method of sampling is that rather than making an attempt to characterize strain in the entire study area, the consideration instead of only rocks showing moderate to intense deformation should represent a maximum strain "grade" for the area and thereby the extreme limits of behavior.

**Limestones.** Limestones were found only within the Ordovician Mazarn and Womble Formations. They are characteristically "marblized" with the degree of recrystallization ranging from samples with ghosts of original lamination preserved to samples showing evidence of multiple periods of recrystallization. The resultant calcite crystals commonly show well-developed deformation lamellae as well as inter- and intra-granular fractures. By analogy with the experimental deformation of calcite rocks (Tobin and Donath, 1971), those from the study area appear to have deformed in a semi-ductile mode and can be classed as showing features of both cataclastic and plastic flow behaviors.

**Chert.** Massive cherts were found in the Ordovician Blakely and Bigfork Formations, and the Devonian-Mississippian Arkansas Novaculite. Unfortunately, they tend to be cryptocrystalline and do not show unambiguous strain features. However, well-formed chalcedony crystals were found in all formations stratigraphically lower than the Arkansas Novaculite. These crystals are commonly sigmoidal or folded and, although they show well-developed undulatory extinction, they are devoid of any deformation lamellae. Intergranular fracturing is totally absent, but some intragranular fractures (macrofractures) are present. These

features indicate that the crystallization history was either syntectonic or that the crystals underwent a syntectonic recrystallization (Durney and Ramsay, 1973). Considering that chalcedony is believed to form by inversion and recrystallization of a previously existing opaline phase (White and Corwin, 1961), the latter view is that taken by the author.

Silicified shales are commonly associated with the chalcedony. They show well-developed alignments of platy minerals producing a slaty cleavage and have shear folds and microlithon structures along the cleavage planes.

**Sandstones.** Rocks of this type are present in all formations exposed within the study area, but are truly common only in the Carboniferous Stanley and Jackfork Formations. The degree of deformation of sandstone samples is intense in outcrop but anomalously lacking in thin section since very little evidence of strain is apparent. Orthoquartzites appear as interlocking mosaics of grains and cement, but where recognizable, rounded grains do not show tectonically-induced shape changes, *e.g.* pressure solution, elongation, flattening, etc. Individual grains show no deformation lamellae and undulatory extinction, although present, is only moderately common. In the majority of samples, grains are usually unfractured and those fractures present do not cross grain boundaries. In a minority of samples, especially those from the Arkansas Novaculite, cataclasis is developed, but fully "demolished" grains are lacking and the distortion of shape for round grains is not great.

Shaley sandstones (sandy shales) similarly lack noticeable strain features. Quartz grains are commonly molded by shales and often have been oriented into the plane of a slaty cleavage. Deformation lamellae are absent and undulatory extinction was found to be uncommon. Grains were characteristically unfractured although macrofractures in hand samples were common.

The deformation of sandstones seems consistent with the expected behavior for sub-greenschist grade rocks (Holland and Lambert, 1969). Intergranular slip is inferred to be the "ductile" deformation mechanism since evidence of plastic flow is absent from all samples studied.

**Shales.** Slaty cleavage is ubiquitous within the study area and is the dominant deformation feature within the shales. The cleavage is produced by the parallel alignment of platy minerals oriented at some angle to the bedding surface. The cleavage most commonly occurs as an axial plane cleavage of flexural flow and shear folds, but was also found associated with boudinage, mini-diapirs, and planes of shear style displacements.

The origin of slaty cleavage has in the past been argued, but it is now generally believed (Ramsay, 1967) that cleavage is produced by the mechanical reorientation of grains, rather than through grain recrystallization. This is consistent with the appearance of cleavage within the study area, since recrystallization at such low temperature

and pressure conditions is doubtful. Furthermore, the observed cleavage geometries seem consistent with the interpretation (Dieterich, 1969) that cleavage occurs as a response to displacements in the direction of tectonic elongation, which implies, in this case, intergranular slip.

**Clastic dikes.** Clastic dikes are common throughout the Eastern Ouachitas. Within the Paron-Fourche SW area, they are recognized within the Ordovician Mazarn, Blakely, and Womble Formations and range in size from microscopic to many meters in length. In thin section, these dikes appear as grains filling fractures in some surrounding rock of different mineralogy. Within individual dikes, flow fabrics, grain alignments and in some cases, grading can be observed. The dikes were found associated with axial plane cleavage, tectonic breccias and fracture sets within folds and are interpreted as being tectonic in origin. Within shale-rich units, associated with clastic dikes were occasionally found features which will be referred to as "plumes." These plumes are most often found at the hinges of folds and are irregular areas (often plume-shaped) where compositional mixing between beds of different mineralogies has occurred.

Clastic dikes that are not synsedimentary are usually interpreted as originating by an injection process and they have been taken to be *prima facie* evidence of abnormal fluid pressures (Maxwell, 1962; Moench, 1966; Powell, 1972a). Although the plumes differ in aspect from the clastic dikes, this compositional homogenization is perhaps a variety of "liquification" and fluid injection (Powell, 1972b).

**Diagenetic features.** Although metamorphic effects are slight within the Eastern Ouachitas, diagenetic changes in rock composition are widespread. Just as for metamorphism, any diagenetic features which are syntectonic provide by geochemical constraints on the processes leading to those features similar constraints upon the environment of deformation. Thus, those features within the Paron-Fourche SW area that are obviously syntectonic seem eminently germane to a discussion of rheology.

Within the Paron-Fourche SW area, there is a major diagenetic division between the rocks of the lower Paleozoic (infrastructure) and Carboniferous (suprastructure). The rocks of the lower Paleozoic commonly have had added to their compositions ferroan calcite, dolomite and chalcedony as diagenetic minerals, whereas the only common diagenetic mineral in Carboniferous rocks is non-ferroan calcite.

Within the lower Paleozoic rocks, carbonate mineralization is present in all units with the possible exception of the lower and middle Arkansas Novaculite members. In beds other than limestones, the ferroan calcite and dolomite occur most commonly as euhedral grains, but anhedral masses occupy the interstices between detrital grains. The amount of carbonate mineralization varies widely; within the Bigfork, these diagenetic minerals are only rarely observed whereas within the Mazarn, they locally comprise

20% of the rock volume. Within the Mazarn and Womble Formations, these two phases appear as replacement minerals after sedimentary calcite. Significantly, they morphologically replace calcite which has already undergone recrystallization and plastic deformation.

Chalcedony was found in all formations stratigraphically below the Arkansas Novaculite but the degree of silicification decreased with stratigraphic depth. The chalcedony was observed as a fracture filling, as an interstitial cement, as layers, and as a replacement mineral after both sedimentary and diagenetic carbonates. As previously stated, much of the chalcedony shows evidence of syntectonic recrystallization.

A detailed diagenetic model is beyond the intent of this paper; however, it seems reasonable to assume, because of the uniformity of distribution and succession, that the diagenetic phases reflect a chain of events affecting the entire pre-Stanley section as a unit rather than separate events affecting the formations individually. This view is perhaps supported by experimental work (Hiltabrand et al., 1973), suggesting that diagenetic carbonates should be rare in pelitic sediments and therefore, that finding the same independent succession of phases is improbable. Likewise, the history of the diagenetic chalcedony is debateable, but since no chalcedony appears above the Arkansas Novaculite, and since the degree of silicification is dependent on depth below the Arkansas Novaculite, the chalcedony and Arkansas Novaculite probably have some genetic relation. This uniformity of diagenetic events is consistent with the succession of phases observed in both the Womble and Mazarn Formations: recrystallization and deformation of limestone; replacement by ferroan calcite and dolomite; replacement by chalcedony. The rheologic importance of this argument is that if some of the dolomite and chalcedony are demonstrably syntectonic in nature, then assuming a uniformity of diagenetic events leads to the conclusion that those events occurred syntectonically.

### Rheologic Model

The observations presented are from too small an area to determine a conclusive rheological model for Ouachita deformation. However, they are indicative and mutually supportive of the hypothesis that water served as the controlling factor of rheology as Maxwell (1962) suggested for low grade metamorphic rocks in the Appalachians. Within the study area, calcite deformed in a semi-ductile manner. This behavior was shown experimentally (Rutter, 1972) to be inconsistent with low temperature and pressure conditions, except in the presence of interstitial water. Chalcedony was found as a phase that underwent syntectonic recrystallization. Although not much is known about the mechanical properties of chalcedony, experimentally, it has been shown (White and Corwin, 1961) that its formation is dependent on the presence of a basic fluid, without which only direct inversion of opal to quartz can occur. Intergranular slip was recognized as the dominant deformation mechanism within shales and sandstones and since the degree of macroscopic ductility produced by

such behavior is a function of the amount of internal friction, the highly ductile deformation features that are observed imply low intergranular friction. It has been shown (Hubbert and Rubey, 1959) that the presence of pore fluids in a body under stress decreases internal friction and promotes intergranular slip. Clastic dikes are a common deformation feature throughout the Eastern Ouachitas and have elsewhere been correlated with abnormal fluid pressures (Maxwell, 1962).

The diagenesis of the Eastern Ouachitas is complicated by a number of phases of apparently syntectonic origin. The unifying factor behind the origin of these phases, however, is the role of fluids in their formation. The origin of dolomite is a matter of argument, but it is generally agreed that saline brines play a significant role. A similar argument has already been stated for chalcedony.

If, as is suggested by the distribution of chalcedony, silica was a mobile species during Ouachita deformation because of some fluid action, then perhaps the Arkansas Novaculite was a water-saturated mass (syntectonically) and solution, pressure solution, and recrystallization may account for its observed ductility. However, no clear evidence of this was recognized within the study area.

Whether or not this hypothesis is reasonable depends to a large extent upon the cause of anomalous fluid pressures during deformation and how such a cause relates to the overall history of the Ouachitas. The unifying factor in Ouachita history is the coincidence of tectonism with geosynclinal development. The rocks of the Ouachitas are divisible into a leptogeosynclinal group and a geosynclinal group with these groups forming the infrastructure and suprastructure, respectively (Viele, 1973). The truly geosynclinal period has been recognized as being concomitant with deformation (Cline, 1970; Viele, 1973) and interdependently, it has been proposed that the deformation was caused by gravity loading (Viele, 1973). Metamorphic (Flawn, 1961) and diagenetic effects also show infrastructure-suprastructure differentiation. Assuming that the above are not just coincidences and that the geosynclinal history of the Ouachitas exercised intimate control over tectonism constrains any cause of anomalous fluid pressures to be a "geosynclinal" process with reasonable probability of being active within the Eastern Ouachitas.

A process which fits this constraint is suggested by the mineralogic differentiation in the Eastern Ouachitas (Weaver, 1961) where the rocks of the infrastructure are dominantly illitic and rocks of the suprastructure are dominantly montmorillonitic. The transition by which montmorillonite becomes illite in a geosynclinal system has been described (Powers, 1967; Perry and Hower, 1972; Schmidt, 1973) as a burial-diagenetic dehydration process, and the salient features are as follows.

Within pelitic sediments, two distinct dehydration phases occur. The first is the expulsion of connate pore fluid through normal overburden compaction, with a

pressure-dependent loss of some interlayer water from expandable clay species. Because of this process, fluid content is reduced to a residual 10% with a potential additional 12% from clay lattice water. This initial dehydration proceeds rapidly with increasing burial for the first thousand meters. However, as porosity decreases with compaction, permeability also decreases, causing the dehydration rate to decrease until the remnant water represents a steady state.

With increasing burial, at depths of approximately 2000 m to 2600 m, depending on the geothermal gradient, mineralogic dehydration begins to not only expel interlayer water, but by lattice collapse to also create new pore space. This dehydration proceeds with increasing burial over a range of approximately 1300 m to 3000 m, again depending on geothermal gradient, and will contribute a total of 10% water by weight to the sediment.

Assuming that the lack of permeability in the overlying non-lattice dewatered rocks would block fluid migration, then, to considerable depth, this 20% by weight pore fluid would remain as a confined fluid. The process is thus one of rock lithification by compaction followed by delithification by lattice dewatering and lattice collapse.

## CONCLUSIONS

The hypothetical rheologic scheme that follows from the above is:

- (1) Before the beginning of Carboniferous geosynclinal sedimentation, the thin leptogeosynclinal/infrastructural rocks underwent compaction and pore space dewatering, making the rocks a rheologically "stiff" mass.
- (2) With the beginning of geosynclinal sedimentation and rapid burial of the infrastructural rocks, lattice dehydration began with the oldest (lowest) rocks first.
- (3) With continuing sedimentation, Carboniferous rocks became compacted and formed a cap for the dewatering lower Paleozoic rocks.
- (4) The water-saturated, delithified infrastructure, through the mechanisms previously described, could deform ductily and did so probably under the influence of gravity loading.

This hypothesis is still highly tentative as it is based on evidence from only a limited study area. However, it does satisfy the constraint that any model seeking to explain Ouachita rheology must satisfy, this being that the cause of deformation, the environment leading to the style of deformation, and the overall history of the Eastern Ouachitas must be intimately related.

## A Semantic Consideration

The above hypothesis is in part an application of new data and a similar tectonic environment to Maxwell's (1962) hypothesis describing the role of a fluid in Appalachian orogenesis. However, his original view was that of a "soft sediment" deformation and whether or not that term can be applied to the Eastern Ouachitas depends on how it is defined. The hypothesis presented herein involves a delithification process which produces a diagenetic "soft sediment." Whether or not this may be included under Maxwell's original term is a somewhat semantic argument

but because of this ambiguity, the author has avoided that term.

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# ROUNDED XENOLITHS IN AN ALKALIC DIKE, GARLAND COUNTY, ARKANSAS

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## INTRODUCTION

An alkalic dike containing rounded xenoliths is exposed by the Union Carbide Vanadium Mine operations in the Potash Sulfur Springs Igneous Complex of Garland County, Arkansas. The complex is located on the southern edge of the Zigzag Mountains Anticlinorium and the northern edge of the Mazarn Basin Synclinorium in the Hot Springs quadrangle, approximately seven miles southeast of the city of Hot Springs on U. S. Highway 270. The massive Arkansas Novaculite and the Hot Springs Sandstone are the resistant units which form the northeast trending Zigzag Mountains--the dominant topographic feature of the area.

The other igneous bodies in the immediate vicinity of Potash Sulfur Springs are the Magnet Cove Igneous Complex which is approximately five miles to the east, the "V" Intrusive approximately one mile to the south, and numerous dikes throughout the area (Fig. 1). The Potash Sulfur Springs Complex crops out over an area of about three-fourths of a square mile and is in contact with the Arkansas Novaculite of Mississippian and Devonian Age, and the Hot Springs Sandstone and the Stanley Shale both of Mississippian Age (Danilchik and Haley, 1964). Mapping by Union Carbide geologists indicates that the intrusion is a ring structure similar to the Magnet Cove Intrusion (Hollingsworth, 1967). The core of the complex is dominantly garnet, pyroxene-rich nepheline syenite, whereas the outer portion is dominantly alkali syenite (Pollock, 1965). Other rock types which occur throughout the complex are nepheline syenite, malignite, fasinite, naujaite, nordjoite, ijolite, and carbonatite. Lamprophyre (fourchite, ouachitite, and monchiquite), tinguaitite, porphyritic phonolite and nepheline syenite dikes cut the intrusion.

## DESCRIPTION OF THE DIKE

An alkalic dike, containing rounded, subangular, "etched" novaculite xenoliths (1-10 cm in diameter) and rounded shale xenoliths (0.1-5.0 cm in diameter) set in a fine-grained white matrix, is exposed in the North Wilson pit of the Union Carbide Vanadium Mine (Fig. 2). This pit is located on the northern edge of the intrusion. Although the dike is irregularly shaped both vertically and horizontally and separates into several distinct portions, the dike is essentially vertical and strikes approximately S. 80° E.

The wall rock is highly fractured, recrystallized novaculite; however, the eastern extension of the dike is in contact with a fenite breccia.

Although the dike appears highly altered in hand specimen, x-ray diffraction analysis indicates that the matrix is dominantly sanadine and some kaolinite. Whole-rock chemical analysis yields a composition approximately that of potassium feldspar which is in agreement with x-ray diffraction data. In addition to the xenoliths, fresh transparent euhedral sanadine phenocrysts (1-4 mm long) and dodecahedra (1-3 mm in diameter) of a fibrous greenish clay mineral, palygorskite (?), are present in the matrix. The palygorskite has probably replaced a sodalite group mineral present as phenocrysts.

Five evenly spaced one-square-yard areas were marked off across the outcrop above the north-south trending berm of the 540-foot level. This is one of the widest exposures of the dike (approximately 90 feet). The number and type of xenoliths in each square were recorded. The shale xenoliths increase in number toward the center of the dike and are distributed symmetrically about the center of the dike (Fig. 3). This distribution of shale xenoliths is similar to the concentration of olivine phenocrysts in the Musk Ox Intrusion (Bhattacharji, 1964) and certain mafic dikes (Hermes, 1964) and is interpreted as the result of "flowage concentration," i. e. the xenoliths have moved toward the center of the dike by the flow of the magma. Therefore, the shale xenoliths must have been obtained from lower stratigraphic units. Flow structures are visible on the outcrop and small elongate xenoliths are locally aligned. A trachytic groundmass is observed in thin section.

The distribution of the novaculite xenoliths is opposite that of the shale xenoliths (Fig. 3). The greatest concentration of novaculite xenoliths is at the contact region of the dike and is approximately symmetrical with respect to the center of the dike. Fewer novaculite xenoliths are present at the north contact possibly because the fractured novaculite of the north contact is well cemented by iron oxides (Fig. 3). Small novaculite xenoliths are sometimes concentrated near the boundaries of larger novaculite xenoliths indicating that the smaller ones have been derived from the larger xenoliths. The novaculite xenolith distribution, thus, implies the incorporation of local wall rock and little flowage concentration.

The most notable feature of the novaculite xenoliths is in their roundness and their characteristically rough, apparently etched surfaces. Local novaculite conglomerate pebbles in comparison are smoother and have percussion marks. A surface texture comparable to that of the

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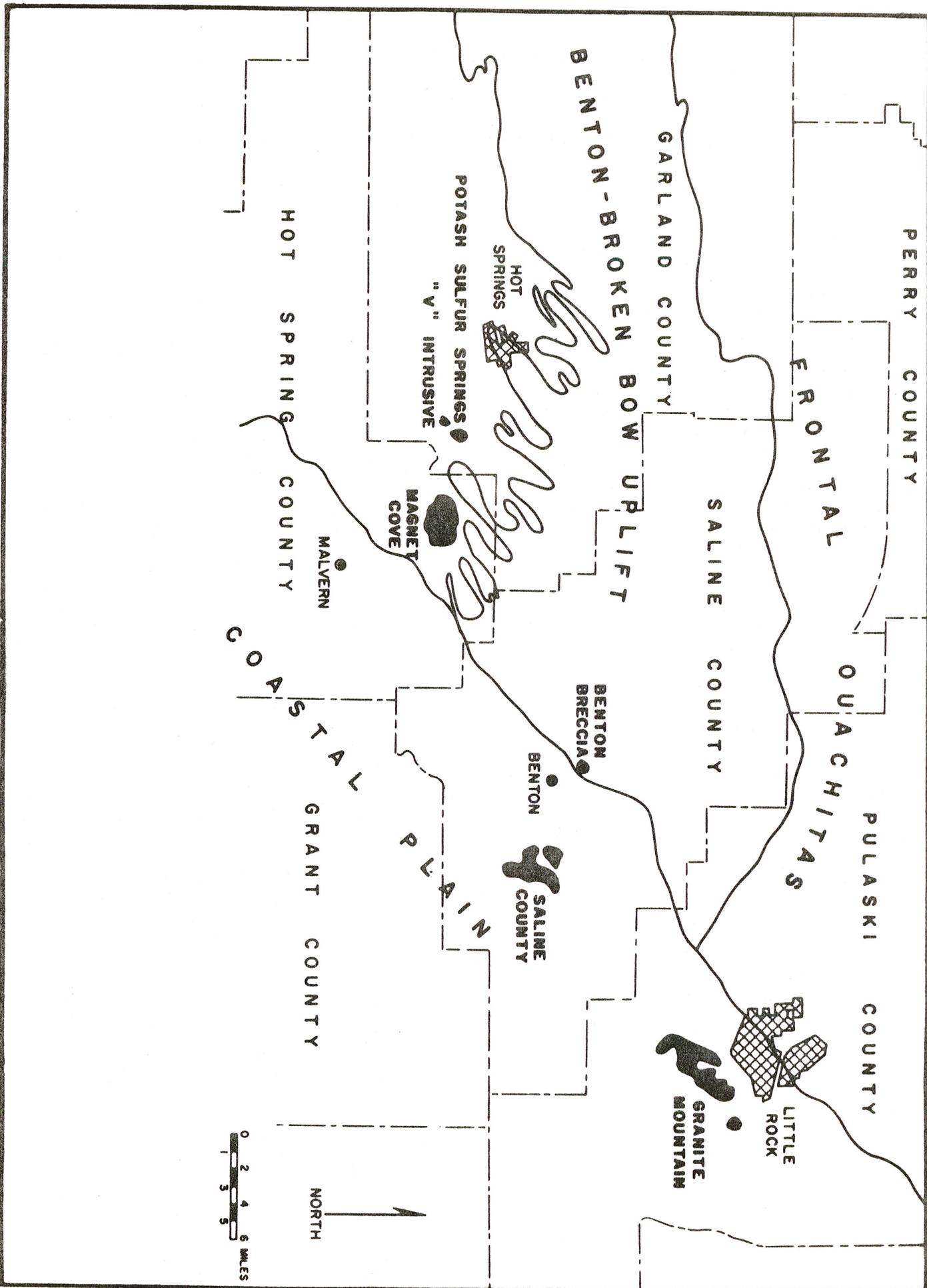


Figure 1. -- Map of exposed alkalic intrusives of central Arkansas.

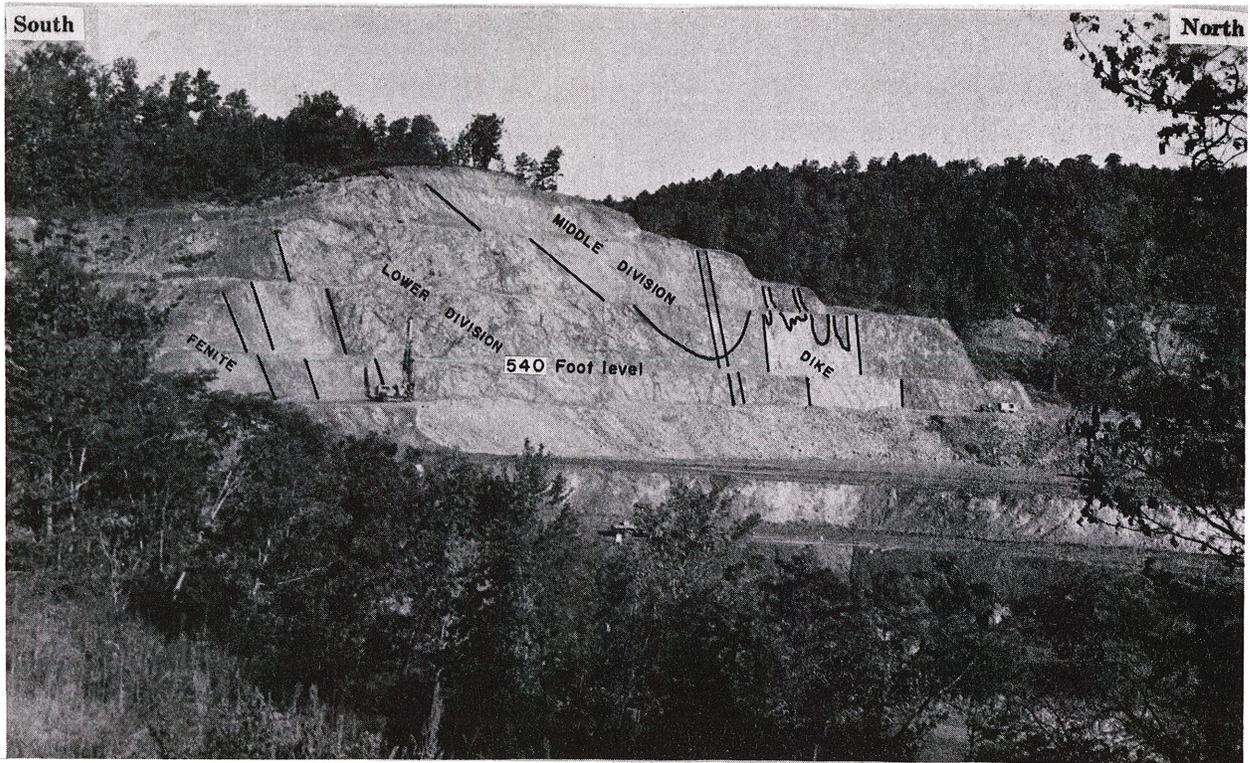


Figure 2. -- North-south highwall of North Wilson Pit. Xenolith-bearing dike in right center of picture (Photograph from Hollingsworth, 1967).

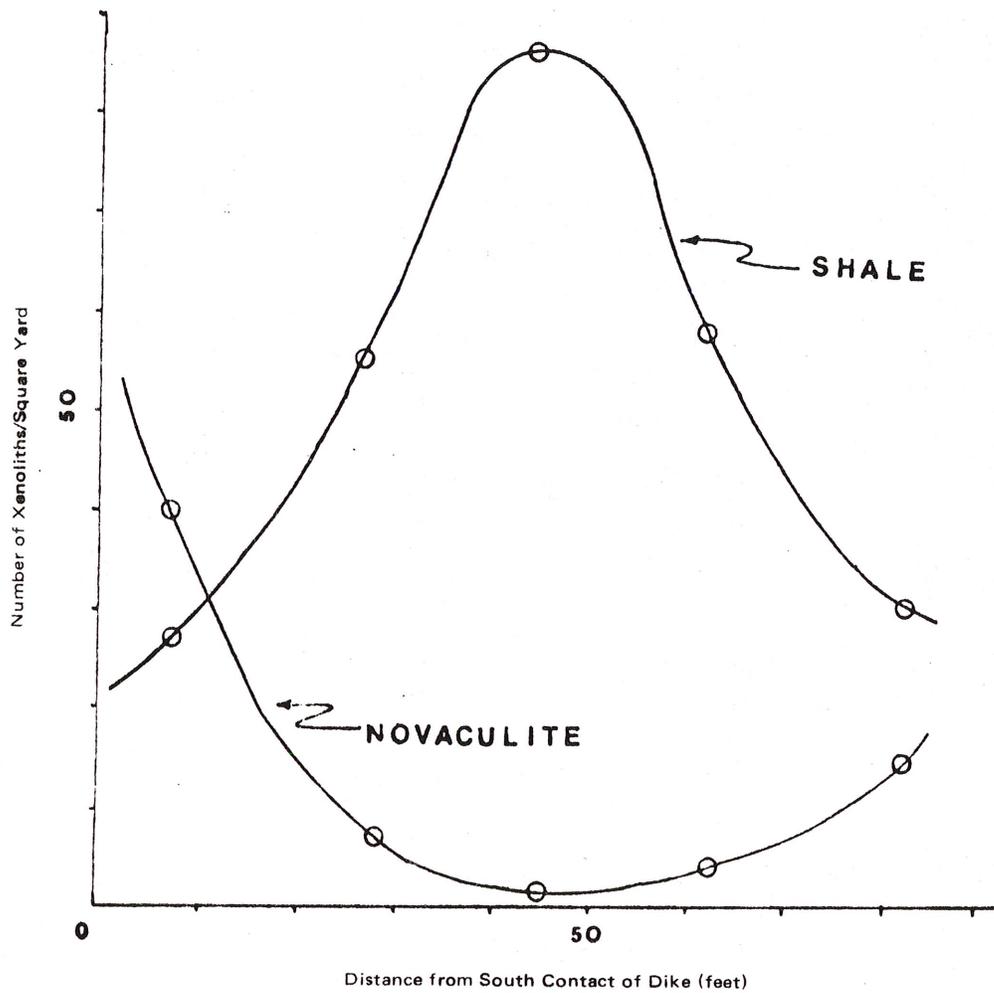


Figure 3. -- Number of dike xenoliths versus distance from south contact of dike.

novaculite xenoliths was produced on a local novaculite conglomerate pebble by exposing it to concentrated hydrofluoric acid for 12 hours. Exposed surfaces of the novaculite contact with the dike are limited due to "limonitic" coatings; however, the few well exposed surfaces have the same rough ("etched") appearance as the novaculite xenoliths.

### CONCLUSIONS

There are several possible causes for the rounding of the xenoliths.

#### 1. Mechanical rounding in a magmatic fluid.

The xenoliths could be rounded due to the abrasion between xenoliths and/or the wall rock; however, no xenoliths were found in contact with each other as might be expected in this situation and no percussion marks have been observed. The distribution of the xenoliths is not that expected using this explanation. The amount of movement necessary for abrasion to cause rounding of the xenoliths should be sufficient to allow flowage concentration of the novaculite xenoliths, as well as shale xenoliths in the central portion of the dike. Repeated fluid pulsations could have moved the xenoliths up and down several times causing abrasion and yet not allowed flowage concentration of the novaculite xenoliths; however, this explanation does not account for the shale xenoliths distribution.

#### 2. Incorporation of stream gravels.

There are no known novaculite gravels in lower stratigraphic units; thus, the novaculite gravels would have to be obtained from higher stratigraphic units. If the novaculite xenoliths were obtained from higher units, their distribution would be expected to be uniform or random across the dike or possibly concentrated in the center of the

dike by magma flow, but *not* concentrated at the dike margins.

#### 3. Chemical rounding by late stage volatile-rich fluids.

Corners and edges of fragments of shale and novaculite would be attacked fastest by a volatile-rich fluid due to their larger surface areas, thus rounding the fragments. Such a fluid would also be capable of etching novaculite, and the small novaculite xenoliths associated with larger ones could perhaps be the result of the chemical action of such a fluid along the fractures of large xenoliths after significant magma movement had ceased. Chemical rounding places no requirements on the source or distribution of the shale and novaculite xenoliths since the rounding takes place *in situ*.

The latter explanation is preferred because: (a) there are no percussion marks on the novaculite xenoliths; (b) the presence of fluorapatite in nearby contact metamorphic rocks, the relatively high fluorine contents in the local rocks and springs, and the carbonatite bodies indicate that a volatile-rich fluid(s) was present during the intrusion of the complex; and (c) it is not inconsistent with the xenolith distribution.

In summary, the shale xenoliths are interpreted as being derived from lower stratigraphic units, concentrated in the central portion of the dike by the flow of a volatile-rich magma and chemically rounded by the fluid(s). The novaculite xenoliths are the result of the incorporation of local novaculite wall rock blocks into the magma during the late stages of the intrusion of the dike (thus voiding flowage concentration) and their subsequent chemical etching and rounding by the fluid(s). Thus, although mechanical abrasion is responsible for the rounding of rock fragments in the vast majority of situations, it appears that in particular cases chemical rounding can be significant and therefore should be considered in determining the geologic history of certain areas.

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Editor's addenda--Steele and others report similar chemically rounded xenoliths in carbonatite intrusions of central Arkansas.