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Petrology of Pennsylvanian Sandstones and Conglomerates of the Ardmore Basin

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TABLE OF CONTENTS

LIST OF ILLUSTRATIONS

LIST OF TABLES

PETROLOGY OF PENNSYLVANIAN SANDSTONES AND CONGLOMERATES OF THE ARDMORE BASIN

Lynn Jacobsen

ABSTRACT

The Pennsylvanian system in the Ardmore Basin comprises as much as 18,000 feet of sediment, of which the greater part is shale, about 20 percent is sandstone, and small amounts are conglomerate and limestone. The major stratigraphic units recognized in the basin are, from the oldest to the youngest, the Goddard shale, Springer formation, Dornick Hills, Deese, Hoxbar, and Pontotoc groups. These sediments are preserved in a narrow syncline between uplifts of the Arbuckle and Wichita Mountains in south-central Oklahoma, and the basin margins are in considerable part high-angle faults of large displacement.

Non-arkosic conglomerates occur at twenty or more stratigraphic levels in the Dornick Hills, Deese, and Hoxbar groups. They are present as wedge-shaped units of rock which thin toward the center of the basin and thicken toward the margins. Chert is by far the most abundant pebble type in the conglomerates, though limestone is locally abundant where the conglomerates are exposed near the borders of the basin. The modal size of the pebbles is about ½ inch, but fragments reach a maximum of 12 inches in diameter. In most occurrences the conglomerates have a quartzose sand matrix which gives them a strongly bimodal size distribution.

The conglomerates are interpreted as fan-shaped basin margin accumulations which were localized by fault-controlled topography. They are significant because they show that deformation of the bordering uplifts was intermittent throughout the period of Pennsylvanian sedimentation. The distribution of the conglomerates suggests that faulting and sharp uplift began on the southern margin of the basin during Dornick Hills deposition but on the north and northeast margins not until after the beginning of Deese sedimentation.

Four relatively well-defined types of sandstone are recognized. One is essentially restricted to the Springer formation, and is characterized by having either quartz alone or quartz and "clay" as the only major constituents, and a tourmaline, zircon, and rutile assemblage of non-opaque accessory minerals.

A second type makes up the sandstones of the Dornick Hills and Deese groups, and much of those of the Hoxbar group. These sandstones are characterized by abundant rock fragments, mostly shale, and a heavy mineral assemblage of garnet, staurolite, chloritoid, chlorite, and clinozoisite, in addition to the ultra-stable minerals.

These two types are similar in texture. Both are very fine to fine grained, and very well sorted. The grains are relatively angular, and have a mean roundness of about 0.40. Bonding is generally weak, but there are examples of bonding by authigenesis of clay minerals, and cementation with calcite, siderite, or secondary quartz.

The third, and least abundant, of the three sandstone types was found only in outcrops of the Hoxbar group. Its distinguishing features are the presence of well-rounded quartz and stable heavy minerals, and abundant detrital limestone fragments. These constituents make up an easily recognized polar type, but most of the Hoxbar sandstones are mixtures, in a considerable range of proportions, of this assemblage and that which characterizes the Deese and Dornick Hills sandstones.

A fourth type is the arkoses of the Pontotoc group. These are tectonic arkoses, and show little modification by chemical weathering and transport. The granitic debris is commonly mixed with sedimentary rock fragments, including limestone, from the earlier Pennsylvanian and pre-Pennsylvanian formations of the Arbuckle Mountains. The arkoses are the distinctive rocks of the Pontotoc group, but Pontotoc sandstones similar in lithology to those of the Deese and Hoxbar groups are probably more common.

The immediate source of the Springer sandstones could not be satisfactorily determined by comparative petrography. The ultimate source was probably a largely metamorphic terrane, and the sandstones have become texturally and mineralogically relatively mature by the largely physical breakdown of their less stable constituents in several cycles of sedimentation.

The post-Springer sandstones are chiefly a mixture of detritus from nearby uplifts, and in a general way record the progressive unroofing of these uplifts. The Deese and Dornick Hills sandstones are largely the product of the erosion of the Springer sandstones; the Hoxbar marks the exposure of the Ordovician orthoquartzites of the Arbuckle Mountains; and the Pontotoc group, the exposure of granite.

Post-Springer sedimentation was ultra-rapid, and as a result the detritus received minimum modification, and for the most part the sediments closely reflect their provenance. Mixing of detritus from more than one uplift or from several exposed lithologies is general. This is most conspicuous in the Hoxbar and Pontotoc sandstones, but all of the sediments may be regarded as a continuous series of increasing complexity with time as a result of exposure of additional areas and lithologies to erosion.

A special instance of mixing is the presence of a group of heavy minerals of metamorphic origin in most of the post-Springer sandstones. These can be shown to have had a source separate from the bulk of the now stable minerals, and are interpreted to indicate a relatively early Pennsylvanian exposure of basement rocks in the Red River Uplift.

The local source, the thick sequence of sediments, the overall similarity of the sequence, and the common occurrence of basin-

margin conglomerates suggest that deformation was essentially continuous during post-Springer sedimentation, and was characterized by abrupt local uplifts.

Although the conspicuous deformation occurred after the deposition of Springer sediments, some structural differentiation of the uplifts had occurred during Springer sedimentation. This is shown by the widespread slumped bedding on the flanks of the large oil field anticlines, and the occurrence of a distinctive facies on and near the crest of these uplifts. The Springer rocks in general show evidence of being deposited in a somewhat stagnant quiet water environment, but on these structural highs the indicated environment is one of strongly agitated water which was oxidizing and alkaline.

The early growth of these uplifts was a critical factor in the petroleum accumulation which occurred in the Springer sandstones in that it caused the sedimentation of a favorable reservoir facies in proximity to a favorable source facies.

INTRODUCTION

Location and Size of Area

The Ardmore basin is essentially an accumulation of Pennsylvanian sediments preserved in what is now a faulted syncline or half-graben. This sequence of rocks is exposed in an area of slightly more than 100 square miles around the city of Ardmore in south-central Oklahoma, but the basin and the same sedimentary facies

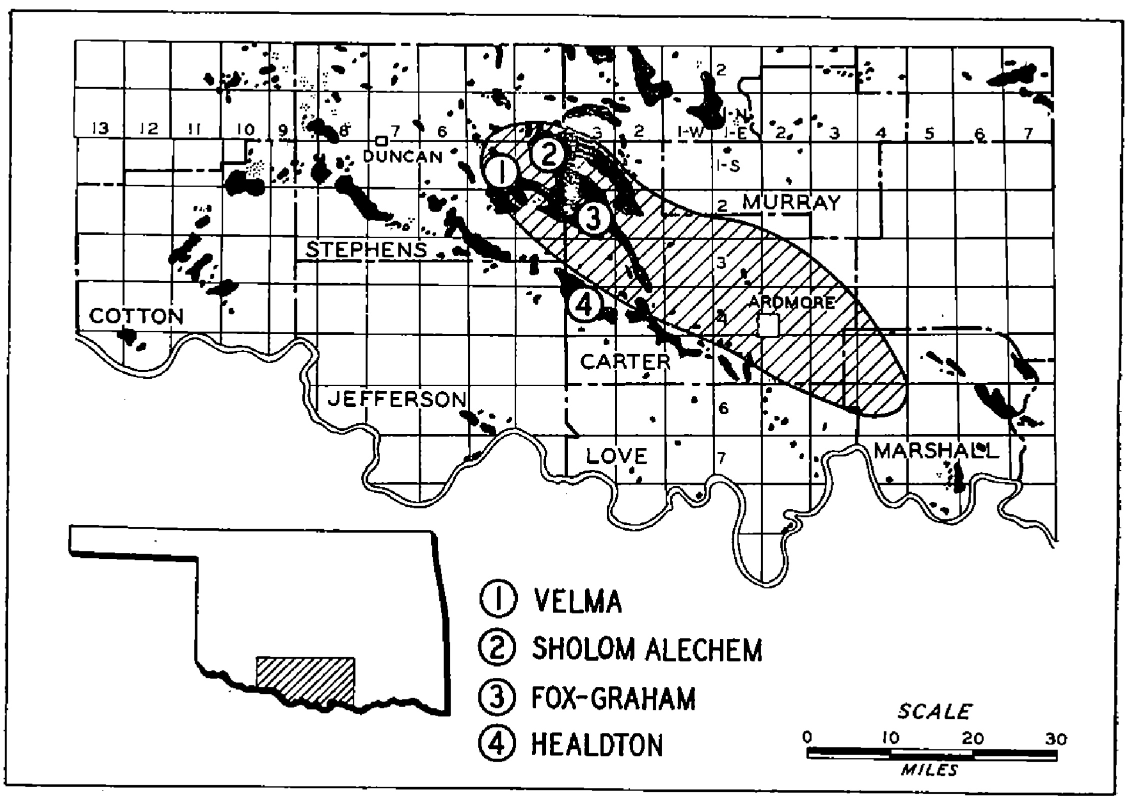


FIGURE 1. Map of southern Oklahoma, showing area studied and location of important oil-producing areas.

continues to the northwest and southeast under a cover of younger rocks. The northwestern part of the basin contains several of the more prolific oil fields in Oklahoma (Figure 1).

The Ardmore basin is not sharply differentiated from the larger Anadarko basin of western Oklahoma, and herein the limit has rather arbitrarily been drawn at the large uplifts which mark the Velma and Sholem Alechem oil fields. Figure 1 shows the area for which the rock descriptions and the general conclusions are intended to apply. This includes approximately 500 square miles.

Maximum thickness of the Pennsylvanian sediments in this area is of the order of 18,000 feet, but the average preserved thickness is less than one-half of the maximum, and is estimated to be about 8,000 feet. It follows then that the volume of Pennsylvanian sediments which it is intended to characterize is approximately 750 cubic miles.

Acknowledgments

This work is the greater part of my doctoral dissertation at The Pennsylvania State College. Completion of the work was materially assisted by the award of the Gulf fellowship in sedimentology and a grant from the Evan Rugh scholarship fund of the School of Mineral Industries of The Pennsylvania State College.

Almost without exception oil companies and individuals within the oil industry cooperated wholeheartedly with the writer. Approximately one-half of the samples studied were from well cores, and for these I am indebted to the following individuals and companies: C. W. Tomlinson, independent oil producer; R. B. Rutledge and Gerald Leukes of the Skelly Oil Company; Erwin Selk of the Pan American Oil Corporation; Arthur Detmar of the Magnolia Petroleum Corporation; Earl Westmoreland and Meryl Williams of the Seaboard Oil Company; John Mayes of The Texas Company; Fred Jones of the Gulf Oil Company; S. R. Woods of the Sun Oil Company; W. Armstrong of the Ohio Oil Company; Curtis Turney of the Stephens Petroleum Corporation; and the Sun Drilling Company.

Dr. C. W. Tomlinson has read the entire manuscript, and has offered a large number of constructive criticisms based on his unequalled familiarity with the Ardmore Basin. Any errors of fact or interpretation which remain are, of course, my responsibility.

I owe a lasting indebtedness to those without whose teaching this work would never have been undertaken. Drs. P. D. Krynine and J. C. Griffiths have given assistance and counsel throughout the study. It will be apparent how much I owe to Dr. Krynine's pioneer work in the synthesis of sedimentary petrography and tectonics, and to his concept of the specific reservoir.

Finally, I wish to acknowledge the assistance of my wife, Eloise Jacobsen, who has given generously of time and encouragement.

Purpose and Scope

This work is concerned primarily with obtaining petrographic evidence which may contribute to a better understanding of late Paleozoic geography of the south-central part of the continent. In practice, this has been a search for petrographic differences within the Ardmore basin sediments, and for similarities between the Ardmore sediments and the rocks of the adjacent uplifts.

The Ardmore sediments are only a small part of the sediments of the region, and it is not to be hoped that the answer to all problems of the region will be answered in this small area. Nevertheless, because of the completeness of the record here and because of the critical regional setting, these sediments offer a particularly inviting subject for detailed study.

Since the study has been primarily a search for criteria useful in paleogeographic reconstruction, it has dealt almost entirely with the detrital rocks, and particularly with the sandstones and conglomerates. These are the rocks which can be expected to yield information as to provenance and are the most rewarding relative to the time and effort expended on them. The shales, although they comprise the bulk of the rocks, and the limestones have been studied only to a minor extent.

Neither have the arkoses of the Pontotoc formation been studied. There is no problem concerning the source of these rocks; they obviously were derived from the granites of the Wichita or Arbuckle Mountains.

Inasmuch as there is little published detailed petrographic information on the rocks which may have been the source of the Ardmore sediments, it has been necessary to extend the study to include them. A reconnaissance petrographic study was made of the rocks of both Arbuckle and Ouachita facies to supplement the available information.

The first field work for the study was done in a short period in the summer of 1949. At that time about sixty samples were collected, and later there were studied in preliminary fashion to guide further sampling. The field work was completed in the summer of 1951. As the Ardmore area has been mapped in considerable detail (Tomlinson, 1929), the field work consisted only of collection of samples and observation of gross sedimentary features.

A total of 550 samples was collected. Of these approximately 100 made up the regional collection, and the remainder were of the Ardmore sediments. Only a small part of these were studied in detail; for example, 110 thin sections were studied, and 89 size analyses and 70 heavy mineral separations were made.

The greater part of this work has been a laboratory study of the collected samples, using ordinary petrographic techniques. Emphasis has been on the study of mineral composition both in thin sections and heavy mineral separates.

The Sampling Problem

The number of samples which have been studied in considerable detail is about 150; this means there is one sample for each five cubic miles of sediment studied. Volumetrically, the proportion of samples to the total amount of sediment is extremely small, but the number of samples necessary to characterize a unit of rock, or any other population, is mainly dependent upon factors other than volume. For example, to determine the average composition of ocean water and river water many more samples would be necessary to characterize the river water although the volume is much less. The number of samples necessary to characterize a population is not dependent on its size, but upon (1) the variability of the population, (2) the precision with which it is desired to characterize the population, and (3) the precision of the measuring technique and the magnitude of experimental errors.

To draw inferences concerning a rock unit from a set of samples it is only necessary that the samples be representative of the whole. The best assurance that samples are representative is that they have been selected on some basis completely independent of the attribute which is to be measured. For example, it is not possible to characterize the grain size of a rock unit by selecting samples on the basis of grain size. To select typical samples one must know all of the rock characteristics.

Proper sampling of sedimentary rocks means that samples must be selected independently of mineral composition and texture. Unfortunately, the attainment of the ideal is not possible; rocks are not exposed fortuitously, and exposures are not independent of texture and mineral composition. The coarser, more quartzose, and better-cemented rocks are necessarily preferentially available for sampling. Similarly cores are not taken at random but because of certain secondary properties, such as oil saturation or rate of penetration, which are related to texture and mineral composition.

A further factor which may constitute a sampling problem is that most of the samples available both from outcrops and well cores come from structurally high areas. There is no necessary relationship between structure and petrography, but it is possible that structural differentiation during sedimentation may have influenced the environment of accumulation. Evidence from the Springer sandstones indicates this is almost certainly true for Springer sedimentation, and it is very likely true for most of the sediments.

In a paleogeographic study the rocks which it is desired to characterize are not those which are now in the basin, but these which were deposited. To the extent that erosion has been selective, the present rocks are not illustrative of the deposited rocks. In the Ardmore basin this is a serious limitation inasmuch as the conglomeratic basin margin accumulations are only fragmentarily preserved. As these are the rocks which could be expected to yield the most information on provenance, their loss obscures the pattern of sedimentation and makes more difficult the general problem of paleogeographic reconstruction.

These limitations on sampling are unavoidable, and they necessarily circumscribe the precision and accuracy of the solution to any geologic problem.

The samples used in this study were selected on the basis of stratigraphic and geographic position; and hence, within the restrictions discussed above, they were essentially random. Further limitations will be brought out as the data are discussed.

REGIONAL SETTING

Structure

The Ardmore basin is located at the southern margin of the central stable region of the United States, and is a part of the deformed area of southern Oklahoma and adjoining states (figure 1). The Arbuckle Mountains form the north and northeast boundary of the basin, and the Criner Hills and buried parts of the Wichita Mountians are the southern margin. The Ouachita Mountains are only a few miles east of the basin.

The Ardmore basin is separated by relatively minor uplifts from the larger Anadarko basin, and together the two form the deep narrow trough which parallels the Wichita Mountains to the north.

The general structure of the Wichita Mountains is a series of west by northwest trending sharp uplifts of Precambrian and lower Paleozoic rocks. These uplifts are separated by basins in which there are thick accumulations of Pennsylvanian and Permian sediments derived in considerable part from the uplifts. The Wichita system is overlapped extensively by these sediments, and it is only a small part of the system which is now exposed. Deformation in the Wichita Mountains involved the granite basement as well as the sediments, and consisted primarily of high-angle faulting and sharp folding.

The Arbuckle Mountains, which form the northern boundary of the Ardmore basin, are a large elongate faulted anticline. Precambrian granite is exposed in three areas on the anticline, and the entire Paleozoic stratigraphic sequence is well exposed on the steeply dipping flanks of the mountains.

The Ouachita Mountains of southeastern Oklahoma and southwestern Arkansas are believed to represent only a small part of a major deformed belt which extends from Mississippi to West Texas. The greater part of the system is buried beneath Coastal Plain sediments, but by means of drill records the Ouachita rock facies have been traced from Oklahoma southeastward around the Llano uplift, and thence westward to connect with the Marathon uplift of West Texas.

The rocks of the exposed Ouachita Mountains are tightly folded and cut by a large number of thrust faults. In Oklahoma the northwest margin of the mountains is marked by a zone of five major thrust faults (Hendricks, *et al.* 1947) which are interpreted as having successively overridden one another to the northwest. The faults all die out as they are traced eastward into Arkansas, and to the south they pass under the cover of Cretaceous sediments. In the interior of the mountains folding is more prominent than faulting; and the rocks are thrown into tight folds, including fan folds (Honess, 1923). Thrust faults which dip to the north are known from the southern part of the exposed Ouachitas (Honess, 1923).

The extension of the Ouachita Mountains beneath the Coastal Plain has been traced by drill cores of steeply dipping rocks of facies similar to those of the Ouachita Mountains. No details of the structure are known; and although the front of this zone is generally thought to be a thrust fault (van der Gracht, 1930), the evidence is not conclusive that this front is not mainly a change in sedimentary facies.

The main uplift of the Wichita Mountains can be dated fairly precisely as occurring in early Pennsylvanian time after the deposition of the Springer formation. Nearly everywhere in southern Oklahoma except the Arbuckle Mountains there is an angular unconformity between Springer and later formations, and the Dornick Hills group along the front of the mountains consists in part of coarse conglomerates containing cobbles of the earlier Paleozoic rocks. In the Anadarko basin arkose is present in the Deese sediments (Wheeler, 1948), indicating that during Deese sedimentation granite had been exposed to erosion locally in the Wichita Mountains. This early Pennsylvanian deformation is known as the Wichita orogeny (van der Gracht, 1930).

The main Arbuckle deformation was in late Pennsylvanian time and is marked by the unconformity below the Pontotoc formation and by the presence of arkosic conglomerates in the Pontotoc. However, local unconformities and conglomerates along the north flank of the Arbuckles (Morgan, 1924) indicate intermittent uplift of this flank occurred throughout middle and late Pennsylvanian time.

Recently Tomlinson (1951) has emphasized that although there are two conspicuous periods of deformation in southern Oklahoma, uplift and deformation was probably continuous throughout Pennsylvanian time.

Diastrophism in the Ouachita Mountains cannot be precisely dated. No major unconformities are found in the Paleozoic sediments which flank the mountains, and dating the uplift is dependent upon recognition of the first sediments which had a source in the Ouachita Mountains. Melton (1930) has argued strongly that there is no evidence that the Ouachita Mountains supplied sediment during Pennsylvanian time; but Tomlinson (1929, 1951) has argued that part of the Ardmore basin sediments were derived from the Ouachita Mountains and, hence, there must have been uplift during Pennsylvanian time.

Stratigraphy

The sedimentary rocks of the region occur as two distinctive facies, generally known as the Ouachita and Arbuckle facies (Table 1). The Ouachita facies is known only from the Ouachita Mountains and from wells drilled in their buried extension beneath the Coastal Plain. The Arbuckle facies crops out in the Arbuckle and Wichita Mountains and is known from wells over most of the Mid-Continent region north and west of the Ouachita Mountains.

The exposed pre-Mississippian rocks of Ouachita facies are relatively thin (3,000 feet) and are marked by the great prominence of bedded chert and the dominance of detrital sediments over limestone and dolomite. The Arbuckle facies, in contrast, is thick (12,000 feet) and consists mostly of carbonate rocks.

The two facies are gradational with one another as shown by several exposures of intermediate rock types in the outer Ouachita Mountains (Hendricks, *et al.* 1947; Harlton, 1953). Although the Ouachita Mountains consist of a number of thrust slices, the movement on individual faults has not been great enough to displace the rock sequences from their depositional order.

TABLE I

Generalized stratigraphy of southern Oklahoma

ARBUCKLE FACIES

System	Group or Formation	Thick. (feet)	Character of the rocks
Perm.	Pontotoc	0-1500	Red shale, arkose, and limestone and arkosic conglomerate
Pennsylvanian	Hoxbar	3500	Gray and brown shale, calcareous sandstone, bioclastic limestone, minor chert conglomerate
Pennsylvanian	Deese	3000-8000	Gray, brown, and red shale, massive sandstone, several chert conglomerates, minor limestone
Pennsylvanian	Dornick Hills	2500 4000	Gray and brown shale, massive sandstone, bioclastic limestone, chert and limestone conglomerate
	Springer	2500-3500	Dark gray carbonaceous shale, thin-bedded to massive sandstone
Miss.	Goddard	2000	Dark gray shale, siderite bands
Miss.	Caney	500	Dark gray shale
Miss.	Sycamore	350	Silty dolomitic limestone
Miss.	Woodford	250-600	Brown and gray-brown chert and siliceous shale
Dev. / Sil.	Hunton	150-300	Limestone and shaly limestone, in part very cherty
Ordovician	Sylvan	300	Gray-green soft shale
Ordovician	Viola	800	Limestone, in part very cherty
Ordovician	Simpson	2500	Limestone and dolomite with several thick (100′+) beds of quartzose sandstones, minor green shale
	Arbuckle	5000-8000	Limestone and dolomite
Camb.	Reagan	100-300	Impure arkosic sandstone

OUACHITA FACIES

Formation	Thick. (feet)	Character of the rocks
Atoka	1500-9500	Gray and gray-green shale with some interbedded graywacke sandstone
Jackfork	3500-6500	Massive hard quartzose sandstone with interbedded gray and gray-green shale
Stanley	3500-6000	Gray and gray-green shale with interbedded graywacke sandstone
Arkansas novaculite	250-900	Upper member — white, calcareous, manganiferous; Middle member—dark gray with interbedded dark shale; Lower member — white, massive
Missouri Mt.	60	Red, green, and black hard shale
Blaylock	800	Very fine grained sandstone, dark sh.
Polk Creek	100-200	Black carbonaceous shale
Bigfork	500-700	Dark gray to black chert, cherty limestone, and shale
Womble	250-1000	Gray to green shales and micaceous and chloritic sandstones
Blakely	0-500	Hard quartzose sandstone
Mazarn	1000	Banded black, green, and red shale, minor limestone and sandstone
Crystal Mountain	500-850	Massive hard quartzose sandstone, conglomeratic at base
Collier Lukfata	?	Dark shale with minor limestone and sandstone

Beginning early in Mississippian time detrital sediments from the south were deposited over all of southern Oklahoma, and the contrast in the two facies was largely lost. The Stanley-Jackfork sequence of the Ouachita Mountains is approximately the coarser and thicker equivalent of the Caney-Springer sequence of the Arbuckle Mountains. These rocks are thick; the Stanley and Jackfork formations may be 12,000 feet or more (Miser, 1934; Hendricks, 1947); and the Caney and Springer formations have a maximum thickness of at least 6,000 feet. The Atoka formation is the youngest unit now cropping out in the Ouachita Mountains, and the Stanley and Jackfork formations are areally the most extensive.

The Caney shale and the Springer formation and equivalent rocks in this region are found only in southern Oklahoma. This Springer-Stanley basin was one of the few areas in the United States where sedimentation was continuous from Mississippian into Pennsylvanian time. The early Pennsylvanian period of restricted seas was short lived, and the Middle and Upper Pennsylvanian sediments progressively onlap northward onto the Mississippian and older formations.

The Middle and Upper Pennsylvanian sediments are thickest (16,000 feet) in the area to the north of the Ouachita Mountains, and there consist almost entirely of shale and sandstone. As these rocks are traced northward, the thickness gradually decreases and limestones become progressively more prominent.

The late Pennsylvanian and early Permian sediments of the region are mostly arkoses derived from the Wichita and Arbuckle Mountains. These are locally overlain unconformably by the Cretaceous sediments of the Coastal Plain.

REGIONAL PETROGRAPHY

Igneous Rocks

Igneous rocks are exposed in each of three mountain systems of southern Oklahoma and, in addition, are known from drilling to be present at shallow depths over a considerably larger area. In the Ouachita Mountains the only igneous rocks are quartz veins and minor sills (and some tuffs), but in the Wichita and Arbuckle uplifts Precambrian granite and other igneous rocks are exposed in large areas.

Granite and subordinate other igneous rocks make up nearly all of the exposed Wichita Mountains and crop out in the crestal portions of the Arbuckle uplift. The total area of outcrop is approximately 300 square miles, but in both the Arbuckle and Wichita areas the granite has been partially buried by sediments derived from the uplifts.

The granite is a pink to red alkali feldspar and quartz rock displaying considerable variation in texture.

In all areas it is markedly low in ferromagnesian minerals. Biotite or hornblende are generally present, but their proportion in few cases is more than 2 or 3 per cent (Hoffman, 1930). In the Wichita Mountain granites the dominant feldspars are perthite and orthoclase (*op. cit.*), but the Arbuckle granites are rich in microcline (Taff, 1928). Plagioclase is in nearly every case present in small amounts, and in local phases of the Arbuckle granite it is abundant enough to make the rock a quartz monzonite.

The range in grain size is about 0.1 mm to 10 mm, with the most common size 2 to 4 mm. For the most part the granites are relatively equigranular, but in the western Arbuckle exposures and locally in the Wichita Mountains the rocks are conspicuously porphyritic.

Gabbroic rocks are exposed over an area of approximately twenty square miles on the northeastern flank of the Wichita Mountains and in several smaller areas within the mountains. They are intruded by the granite and are apparently older.

The gabbro is a dark gray rather coarsely crystalline rock. Labradorite is the principal constituent in all occurrences, and the proportion of labradorite ranges from 50 to 98 per cent. Diallage is the chief ferromagnesian mineral; hornblende and biotite occur as minor constituents (Hoffman, 1930).

Dikes of diabase are common throughout the area of granite exposures in the Arbuckle Mountains, and also occur in a few places in the Wichita Mountains. The dike rocks are fine- to medium-grained and distinctly ophitic. Their composition is uniform in both areas; labradorite and augite are the only abundant constituents, but magnetite and biotite are everywhere present.

Quartz veins are abundant in the Ouachita Mountains (Miser, 1944), and locally in the central part of the mountains residual vein quartz literally covers the ground. This is the white massive type of quartz commonly known as "bull" quartz, but the quartz also occurs as small scale veins in nearly all the early Paleozoic sediments.

Metamorphic Rocks

High rank metamorphic rocks.—The only high or moderate rank metamorphic rock exposed in the general region is the Meers quartzite, which is known from a small area in the eastern part of the Wichita Mountains (Hoffman, 1930; Merritt, 1948). This rock is fine- to medium-grained, and ranges from a clean white quartzite to a moderately feldspathic quartzite.

Metamorphic rocks have been encountered as basement in at least one well on the north flank of the Wichita trend (Carter No. 1 Williford, sec. 24, T. 2 N., R. 9 W.), and are present directly beneath Upper Deese or Hoxbar sediments along the Red River Uplift (Flawn, 1956).

Low rank metamorphic rocks.—In the central part of the exposed Ouachita Mountains and in the subsurface continuation of the Ouachita belt the lower Paleozoic sediments have been subjected to moderate metamorphism. The degree of metamorphism does not anywhere reach above the biotite zone, and in the greater part of the area the sediments are essentially unaltered (Goldstein and Reno, 1952).

Of the more intensively metamorphosed rocks the chief effects are the development of wavy bedding or false cleavage in the shales, and local cataclastic deformation of the quartz grains of the sandstones.

According to Goldstein and Reno (1952) the argillaceous material of the shales and of the sandstone matrix has to a considerable extent been reconstituted to sericite and chlorite.

The quartzites which occur in the Crystal Mountain, Blakely, and Jackfork formations are sedimentary quartzites. They are sandstones well-cemented by authigenic silica, and they show few cataclastic effects or formation of new minerals.

Sedimentary Rocks

The general region of southern Oklahoma displays a great variety of sediments in which almost every textural and compositional variety with the exception of evaporite and high rank graywacke is represented. The overall pattern is one of progressive northwestward spread of detrital sediments during the Paleozoic era; this is accompanied by a change in the sandstones from orthoquartzites in the lower Paleozoic to graywackes in the middle Paleozoic, and to arkose in the upper Paleozoic.

In the rocks of lower Paleozoic age the detrital sediments are confined largely to the Ouachita belt; but by late Mississippian time the detritals had spread to central Oklahoma (Selk, 1948); and by the end of the Pennsylvanian period, to western Kansas (Maher, 1947).

The older sandstones of the region were possible source rocks for the Ardmore basin sandstones and have been studied in considerable detail; the other sediments were considered only incidentally.

Graywackes.—The chief regional occurrence of graywackes is in the Stanley and Atoka formations of the Ouachita Mountains. Apparently all of the sandstones in the Stanley formation are typical graywacke, but some of the Atoka sandstones approach being orthoquartzites. Very fine-grained graywacke sandstones or siltstones occur in the Blaylock and Womble formations; but as they are finer-grained than the Ardmore sediments, they are not, in their exposed facies at least, a possible source.

The graywackes consist of 50 to 70 per cent quartz with the balance mostly rock fragments, feldspar, and a clay-mica matrix. The quartz consists predominantly of angular grains of nondescript quartz, but in the Stanley graywackes there is an admixture of well-rounded grains or pebbles of vein or pegmatitic quartz (figure 2). The rock fragments include slate, phyllite, mica schists (figure 3), fine-grained igneous rock (andesite ?), and chert. The feldspar is nearly all plagioclase, apparently in the oligoclase-andesine range.

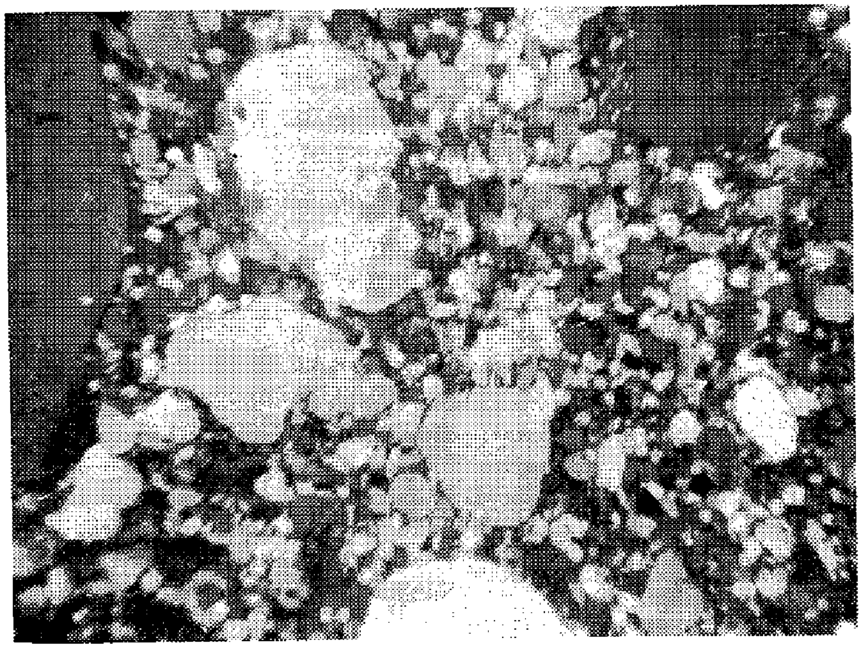


FIGURE 2. Rounded vein quartz fragments in Stanley sandstone (St-508). Crossed nicols.

FIGURE 3. Mica schist pebble in Stanley sandstone (St-510).

Heavy minerals of the graywackes are dominated by garnet, but include tourmaline, zircon, rutile, apatite, chlorite, chloritoid, staurolite, and the usual opaque minerals. Bokman (1953) from his more complete collection also reported sphene, epidote, sillimanite, hornblende, and hypersthene. However, garnet, tourmaline, zircon, and rutile are the only minerals which are present in more than trace amounts. Much of the garnet is intergrown with chlorite, and is possibly being replaced by the chlorite.

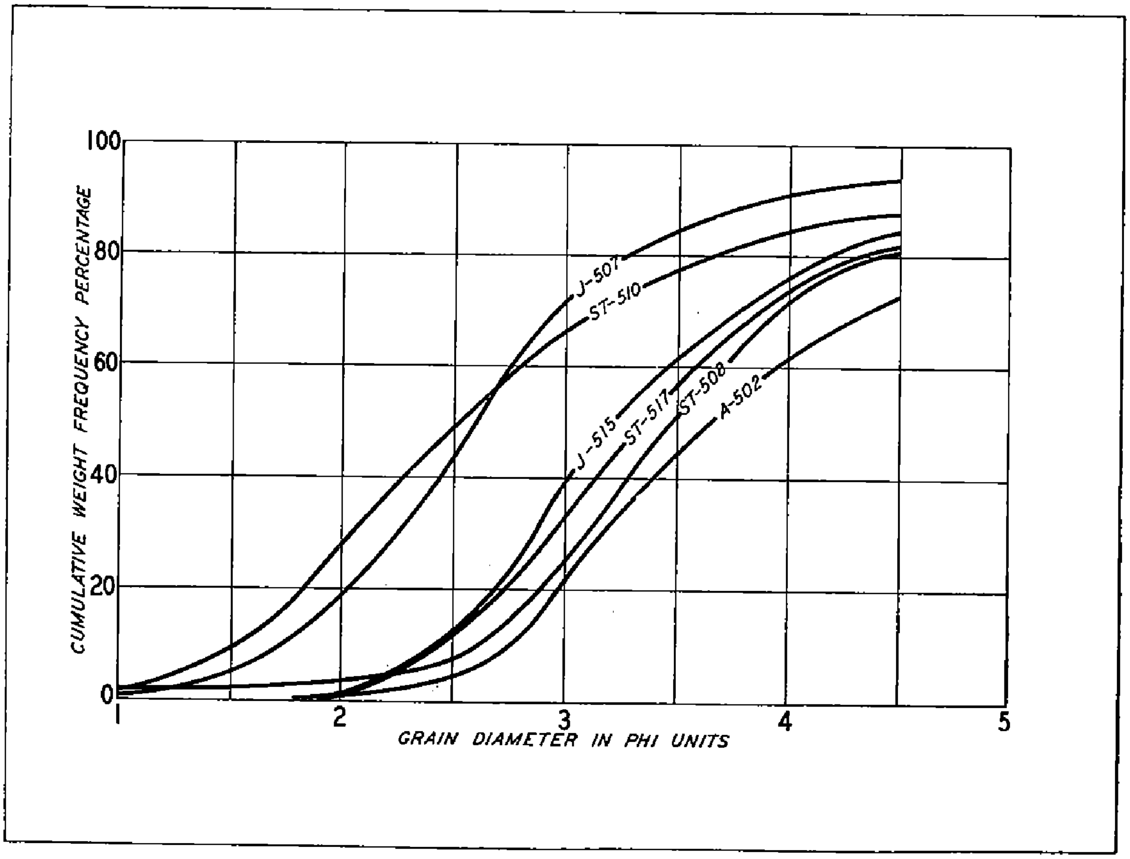


FIGURE 4. Cumulative size frequency distribution of sandstones from the Ouachita Mountains.

Grain size distributions for several Stanley graywackes are shown as cumulative curves in figure 4. These were determined by sieving, and because of breaking of rock fragments the distributions are probably shifted to the fine side. However, from field observations it is thought these distributions are reasonably representative. Nearly all of the Stanley sands are fine-grained or very fine-grained. Except for the vein quartz granules or pebbles, which constitute a separate population, there is little material in the Stanley coarser than one-half millimeter. The sands are all poorly sorted, with sorting coefficients ($PD\phi$) larger than 1.5. The distributions differ from those of Ardmore sediments particularly in the very much poorer sorting of the central part of the distributions.

The nature of the source area of the Stanley graywackes is indicated by the rock fragments which are found in the sandstones. Low to moderate rank metamorphic rocks are most common, and these are accompanied by the vein and pegmatitic quartz and by fragments of igneous flows. The suggested provenance is not essentially different from many parts of the Piedmont province of the eastern United States. As there is no such area to the west, north, or northeast of the Ouachita Mountains, the source area must have been to the south. This is the area which Miser (1921) has referred to as Llanoria. Probably this land area was the geomorphic product of an early phase in the destruction of a a geosyncline which had accumulated sediment during lower Paleozoic time.

Protoquartzites.—The sandstones of the Jackfork formation are quartzose rocks of the type Krynine (1951) has called protoquartzites. These are rocks which in terms of composition are part of the continuum between graywackes and orthoquartzites, but are nearer the orthoquartzite end of the series. The Jackfork sandstones contain 80 to 95 per cent quartz; and the balance is plagioclase feldspar, metamorphic rock fragments, and a clay-mica matrix. Silica cement in the form of grain overgrowths is common (figure 5), and the Jackfork sandstones are characteristically well-bonded and resistant to erosion.

Heavy minerals in the Jackfork are predominantly the ultrastable zircon, tourmaline, and rutile; but trace amounts of garnet, apatite, and staurolite also occur.

The grain size of the Jackfork sandstone is slightly coarser than that of the Stanley sandstone (figure 4), but this is at least partly to be accounted for by the presence of more clay and breakable rock fragments in the Stanley samples. Probably the size of the quartz grains in the two formations is not significantly different.

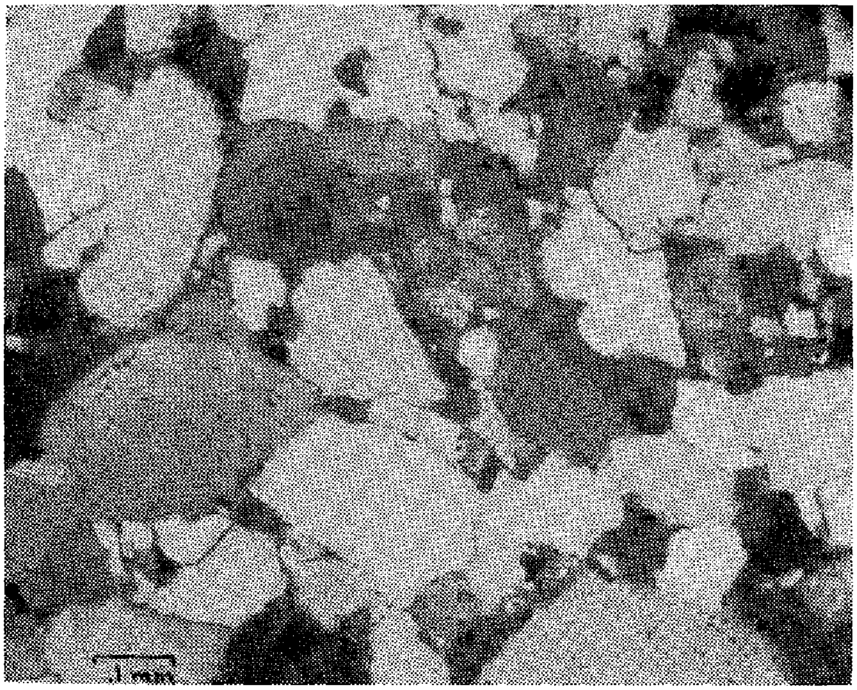


FIGURE 5. Jackfork sandstone (J-507). Crossed nicols.

The Jackfork sandstones differ chiefly from the Stanley sandstones in their lesser content of relatively unstable detrital grains. From their texture and mineral composition it seems likely that the source of the Jackfork sandstones was the Stanley formation or similar rocks. Possibly the area of the Ouachita Mountains is only the northern edge of the Stanley basin of deposition, and the sediments deposited in the southern part of the basin were uplifted as the orogeny which caused the Stanley sedimentation progressively spread northward.

Orthoquartzites.—Second-cycle orthoquartzites (Krynine, 1942) occur in the Simpson group of the Arbuckle and Wichita Mountains, and the Crystal Mountain and Blakely formations of the Ouachita Mountains.[1] These are petrographically similar; variation is relatively slight; and they can be characterized as follows:

(1) Quartz is the only major constituent. In all samples seen it constitutes more than 95 per cent of the detrital material; and in most samples it is 98 to 99 per cent.

[1] These sandstones are here considered to be second cycle orthoquartzites because they occur in the same part of the stratigraphic section as the St. Peter sandstone, and are generally similar to it. The simplest paleogeographic interpretation of these sandstones is that they are reworked Cambrian sandstones, but there is no petrographic evidence available, as far as the writer knows, which would in itself prove that this is so.

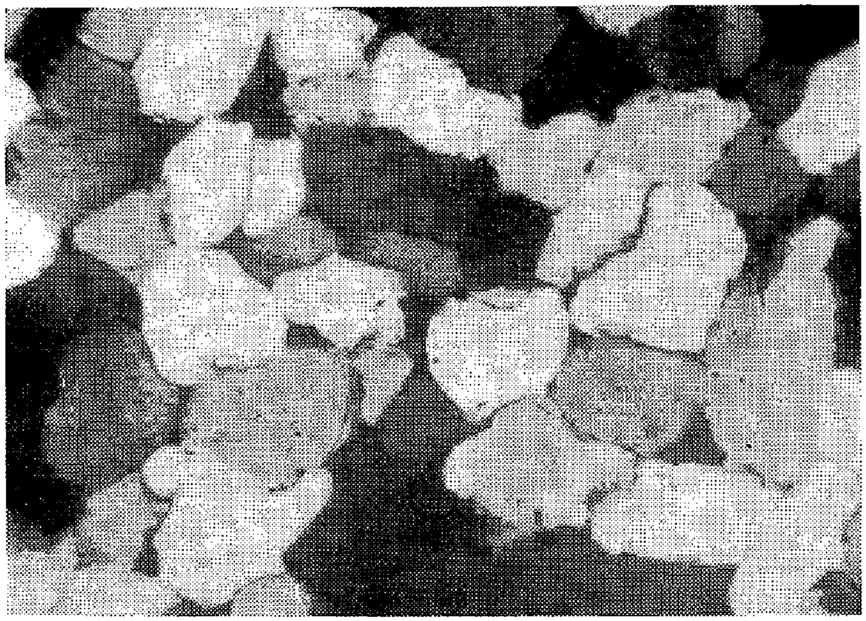


FIGURE 6. Simpson orthoquartzite (Si-246). Crossed nicols.

(2) Silica cement is present nearly everywhere, but the amount ranges widely. In a number of oil reservoirs and also commonly on the outcrop areas the rocks are friable and unbonded, in other occurrences the pores are filled with silica and the rock is tightly bonded. In most instances the silica occurs in the usual form of overgrowths on the quartz grains (figure 7), but in some occurrences of the Crystal Mountain

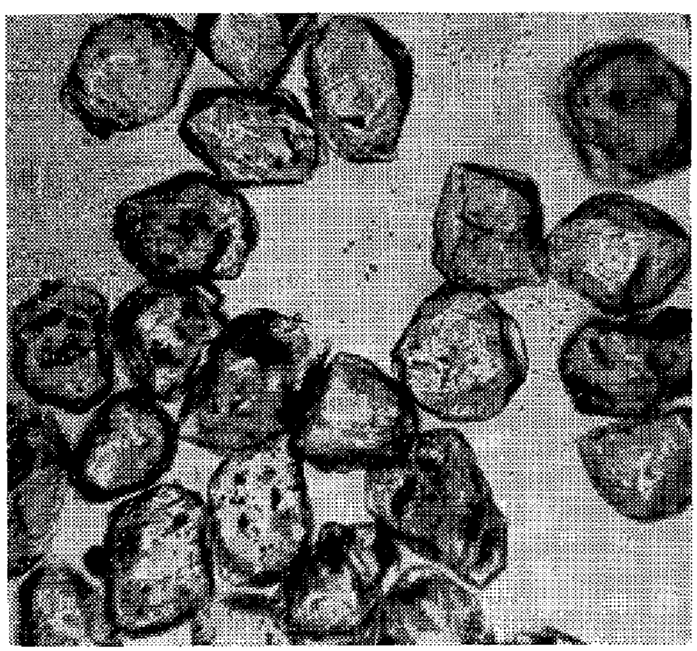
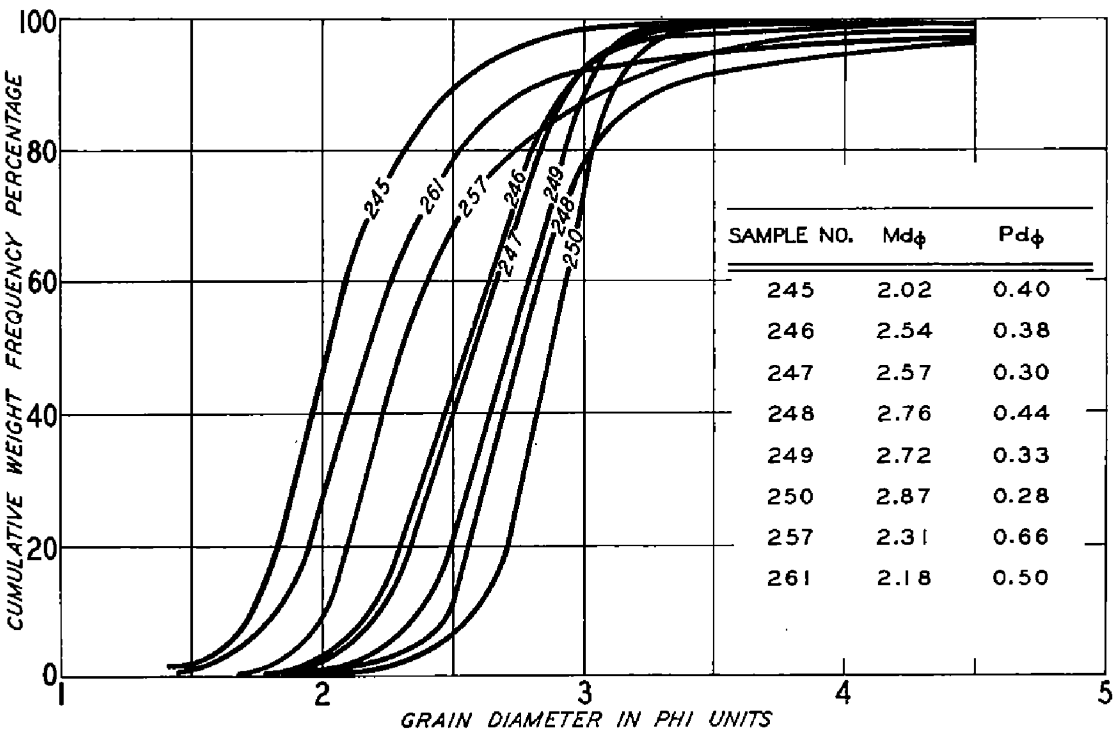


FIGURE 7. Quartz grains with overgrowths, from a Simpson orthoquartzite (Si-261). Plane polarized light.

sandstone the silica cement on adjoining grains is intergrown along such strongly sutured contacts that it appears to be microcrystalline.

(3) Heavy minerals occur in small amounts (less than 0.1 per cent in any size fraction) and few species. All of the detrital heavy minerals are ultra-stable; and the only major ones are tourmaline, zircon, and rutile. A few grains of spinel, both picotite and ceylonite, and rarely garnet are also present. The opaque minerals are less abundant than the non-opaque, and ilmenite is the only abundant one. The authigenic minerals include anatase and wurtzite.

(4) Grain size distributions are characterized by extremely good sorting and the virtual absence of silt or clay sized particles (figure 8). Most of the sands are fine-grained (have a median by sieving between 2 and 3 phi).



SAMPLE NO.	Mdϕ	Pdϕ
245	2.02	0.40
246	2.54	0.38
247	2.57	0.30
248	2.76	0.44
249	2.72	0.33
250	2.87	0.28
257	2.31	0.66
261	2.18	0.50

FIGURE 8. Cumulative size frequency distributions and summary statistics of Simpson sandstones.

(5) The detrital grains are characterized by a high degree of rounding (figures 9 and 10). For the quartz grains the rounding is usually modified by overgrowths (figure 7), but the rounded outline is apparent beneath the overgrowth.

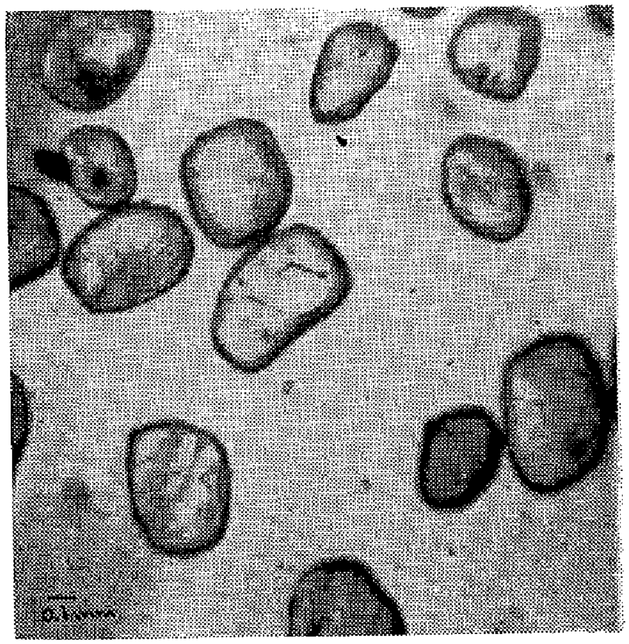


FIGURE 9. Rounded quartz grains from a Simpson sandstone. Plane polarized light.

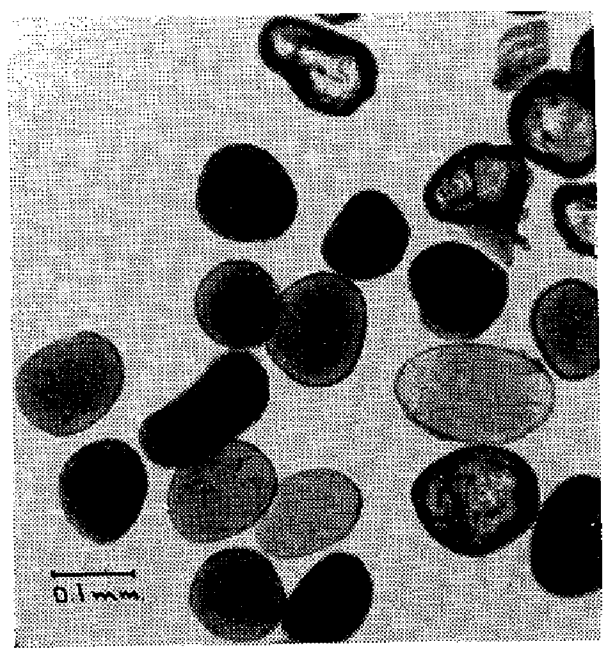


FIGURE 10. Heavy minerals from a Simpson sandstone (Si-245). Plane polarized light.

STRATIGRAPHY[1]

General Description

Approximately 18,000 feet of Pennsylvanian sediments, representing all or nearly all of the period, are preserved and exposed in the Ardmore Basin. The completeness and thickness of the Pennsylvanian section are unique in North America.

Composite surface and subsurface sections for the outcrop area and the buried northwestern part of the basin are given in figures 11 and 12). The surface section is modified slightly from that of the Ardmore Geological Society (1957), and the subsurface section was obtained through the courtesy of the Schlumberger Well Surveying Corporation.

In the Ardmore Basin sedimentation was apparently uninterrupted in late Mississippian and early Pennsylvanian time. This was a period of gradually increasing detrital sedimentation, and the sequence is completely transitional from the limy shales of the Caney to the non-limy ones of the Goddard and progressing to the abundantly sand-bearing Springer formation.

The first abrupt change in the sequence occurs with the beginning of Dornick Hills sedimentation when coarse detritus of local origin is introduced for the first time. The Dornick Hills, Deese, and Hoxbar groups consist of alternating shale, limestone, sandstone, and local conglomerates. The proportions of the different lithologies vary in the several groups, but in general they are similar. However, the Pontotoc group, which overlies the Hoxbar with sharp angular unconformity, consists chiefly of arkose and red shale, and thus makes the second abrupt change.

Coal is known at two horizons in the sequence, one in the Big Branch formation and the other in the Hoxbar; but these deposits are local, and the section is overwhelmingly marine. Fossils in the Springer are sparse and include plant impressions, but the Springer carries a long ranging macrofauna and has also yielded locally a rich marine fauna which Elias (1950) has described as transitional between known Mississippian and Pennsylvanian types. The Dornick Hills, Deese, and Hoxbar groups are richly fossiliferous; and the varied fauna includes brachiopods, pelecypods, gastropods, crinoids, fusulinids, and Bryozoa.

[1] In large part adapted from Tomlinson (1929, 1951).

Goddard Shale

The Goddard is dark gray shale with abundant siderite partings and concretions, and a few thin sandstone beds. On the surface the Goddard is approximately 2,000 feet thick, but in wells in the northwestern part of the basin it is less than 1,500 feet.

The abundant siderite and the dark color of the shale indicate deposition in a reducing environment and probably one of restricted circulation.

In a few localities a near normal bottom dwelling fauna has been found (Elias, 1956; 1957) which suggests that there was, at least in small areas and for short times, an open agitated marine environment. However, fossils are rare, and throughout most of the formation consist of conodonts and scattered goniatites as might be expected in an area of somewhat stagnant bottom conditions.

Conventionally the Mississippian — Pennsylvanian boundary has been considered to be at the base of the Goddard shale, but the recent paleontologic work by Elias (1956) suggests that most if not all of the Goddard is more properly classified as Mississippian.

Springer Formation

Conformably over the Goddard shale is the Springer formation, which consists of approximately 2,000 feet of dark gray shale similar to that of the Goddard, but marked by the presence of four to six prominent sandstone members.

The Springer has been widely accepted as being of early Pennsylvanian age, and has been used as a type section for the Springeran Stage by Moore and Thompson (1949) in their revision of the Pennsylvanian system. However, no unequivocal paleontologic or stratigraphic evidence as to the age of the Springer is now known. Many Oklahoma geologists now believe the Springer is more logically classified as Mississippian than Pennsylvanian, and that as subsurface information becomes available the Springer will be found to grade into Upper Mississippian shales and limestones on the north flank of the Anadarko Basin.

The sandstone zones of the Springer are named, and include, from bottom to top, the Rod Club, Overbrook, and Lake Ardmore sandstones. The Primrose sandstone, about 500 feet above the Lake Ardmore, has been variously classified as Springer and as Dornick Hills, but is now generally considered to be the basal

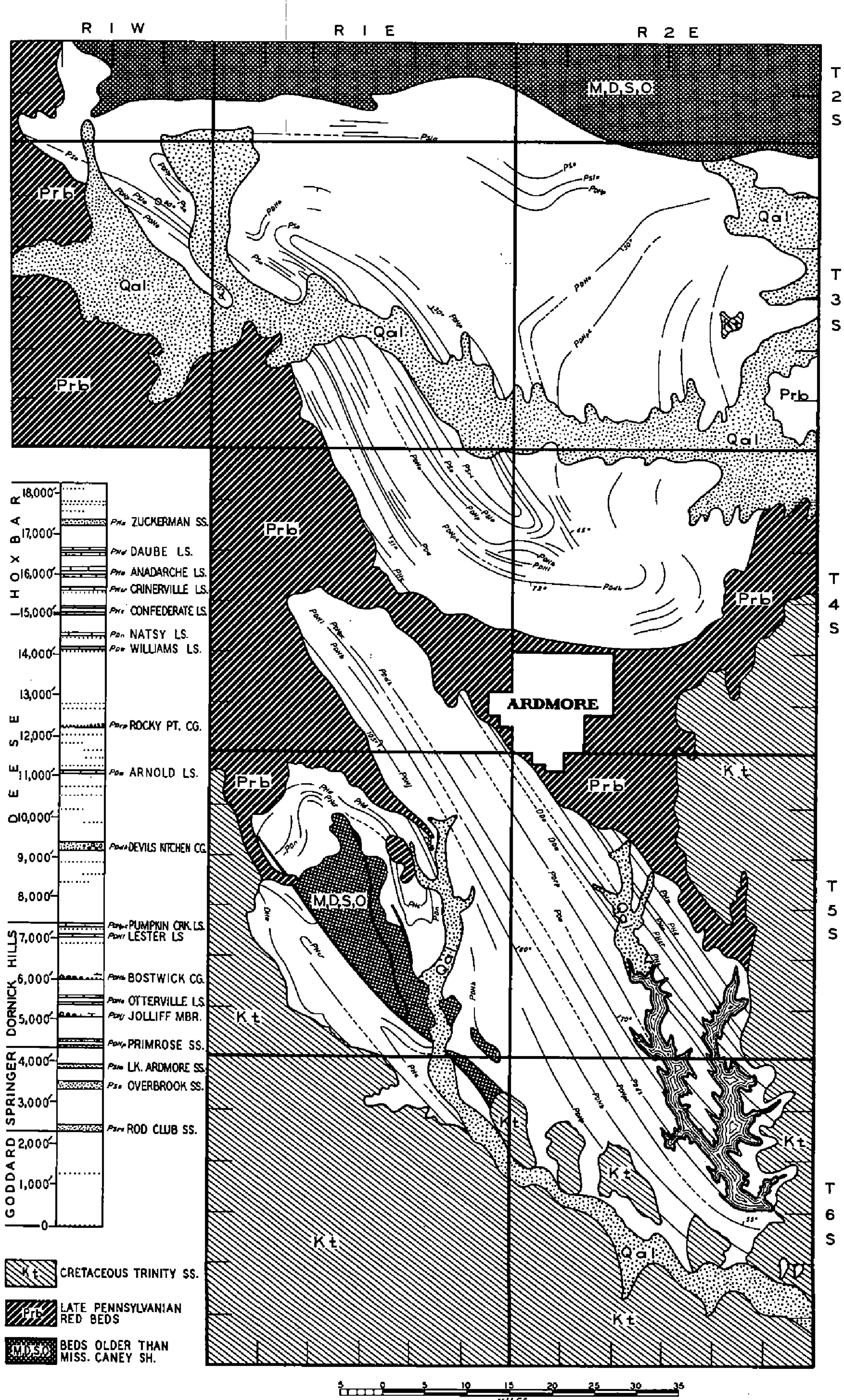


FIGURE 11. Outcrop map and stratigraphic section of the Pennsylvanian system in the Ardmore Basin (adapted from Tomlinson and guidebooks of the Ardmore Geological Society).

member of the Golf Course formation of the Dornick Hills group. All the sandstones are very fine-grained, but otherwise they are quite variable both between members and within themselves. None is a continuous blanket sandstone, and only on the Caddo and South Woodford anticlines northwest of Ardmore are all four exposed.

In this area the Rod Club consists of one to three beds of massive sandstone 15 to 25 feet thick and several thin beds distributed through approximately 300 feet of shale. The sandstones are gray-green or brown, very fine-grained, hard, and with very minor shale partings.

The Overbrook is 100 to 150 feet thick in the same area. It is characteristically light gray and slabby to thin-bedded though varying locally to massive. Some beds are very shaly, but no major shale break interrupts the sequence of sandstone. The Overbrook is asphaltic on the south flank of the Woodford anticline and has been quarried.

About 500 feet above the Overbrook is the Lake Ardmore sandstone, which is similar to the Overbrook except that it is only 15 to 20 feet thick and is commonly brown rather than gray. A thin limestone, the Target member (Bennison, 1954), occurs locally in the lower part of the Lake Ardmore sandstone. It is the only limestone known in the exposed Springer, although the sandstones are in a few places calcareous, and it has been one of the few Springer localities from which a normal bottom dwelling fauna has been collected.

In the northwestern part of the Ardmore Basin Springer sands are prolific oil reservoirs and constitute the chief producing beds of such major fields as Velma and Sholom Alechem and a number of smaller pools in northwestern Carter County and southeastern Stephens County.

In a general way the Springer in subsurface is similar to that of outcrop areas. It is 1,500 to 2,000 feet thick where most completely preserved, and is made up of 75 to 85 per cent uniform dark gray shale with thin sideritic bands. Five sandstone members with an aggregate thickness of 300 to 500 feet are generally recognized. Names in general usage for these are, from top to bottom, the Markham, Aldridge, Humphreys, Sims, and Goodwin sands [Elias (1956, p. 85) suggests the Goodwin is in the upper part of

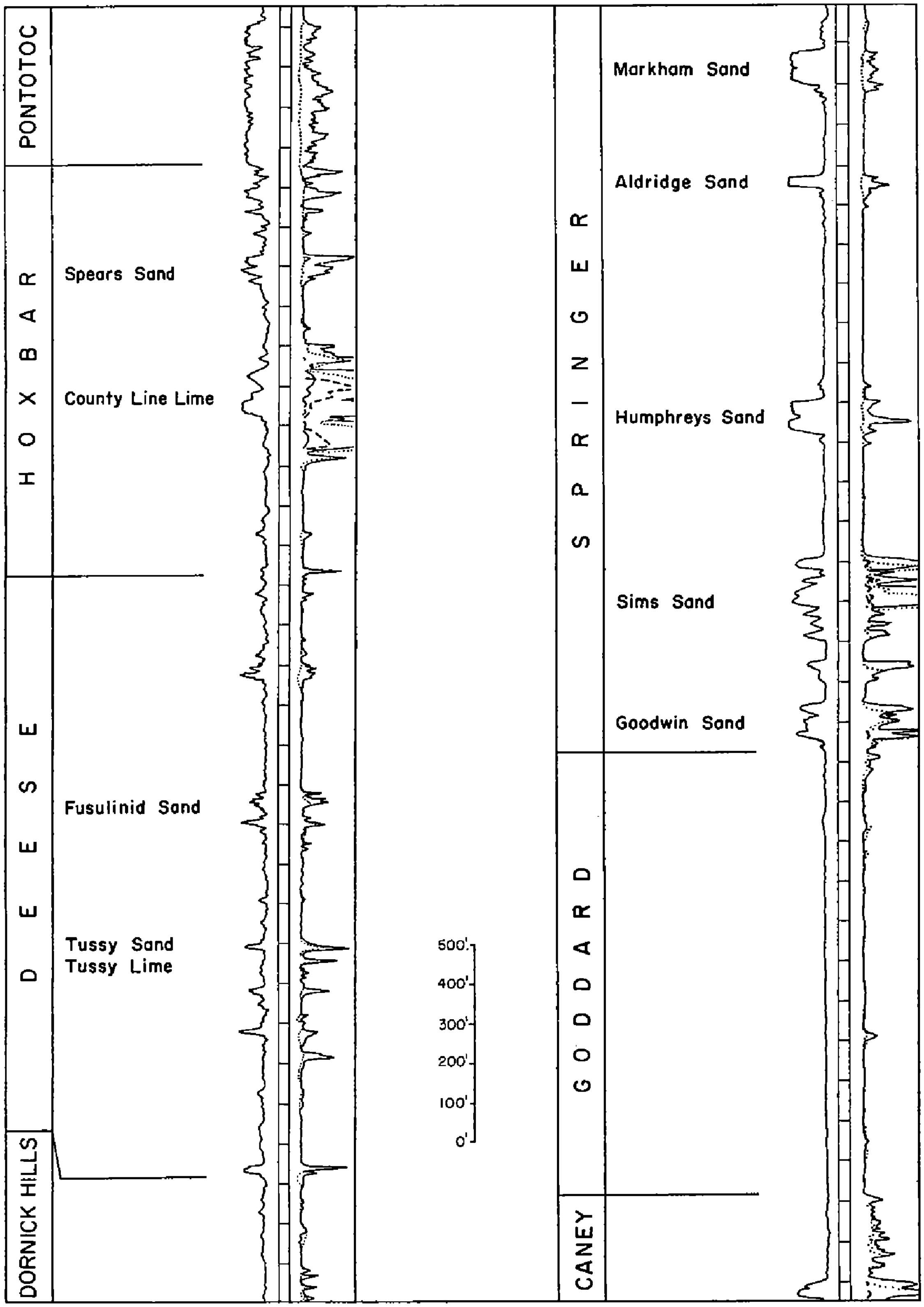


FIGURE 12. Composite electric log stratigraphic section of the Ardmore Basin.

the Goddard shale]. They are quite consistently present over an area of about eight townships where close well spacing gives good control, but they cannot be satisfactorily correlated with the surface units nor with the Springer sandstones which are known farther north in Grady County.

A thin limestone, bearing lenses of black oolites, is generally present slightly above the Humphreys sandstone.

On all of the major uplifts known from wells the Springer is partially or completely truncated, and is overlain either by upper Dornick Hills beds or by Deese beds in an onlapping relationship. The Springer generally thins to the northwest, and is not present very far north on the north flank of the Anadarko Basin. It thickens to the southeast and apparently becomes much thicker south and east of the area studied here.

Dornick Hills Group

The Dornick Hills group is in its bulk characteristics lithologically distinguished from the Springer by a change from dark gray to light brownish gray shale, by the presence of coarse conglomerates, and by the preponderance of limestone over sandstone. In the outcrop area it is 1,500 to 4,000 feet thick. Three formations are recognized; from bottom to top, these are the Golf Course, Lake Murray, and Big Branch formations.

At the base of the Dornick Hills group is the Primrose sandstone. The Primrose consists of very shaly calcareous thin-bedded sandstone interbedded with black shale in a zone 150 to 200 feet thick.

Several hundred feet above the Primrose is the Jolliff limestone and limestone conglomerate. In its type locality in section 24, T. 5 S., R. 1 E., the Jolliff is a limestone bed four feet thick; but both north and south coarse conglomerates are present, and the aggregate thickness of limestone, conglomerate, and shale is as much as 500 feet. Table 2 is the Jolliff section measured at the excellent exposure on Highway No. 70 in sec. 28, T. 4 S., R 1 N. The Jolliff is found only along the east flank of the Overbrook anticline and on the southeast nose of the Caddo anticline. To the west it has been removed by erosion, and to the north it probably grades into shale.

A few thin (less than 6 inches) sandy zones are present in the Jolliff; the sand is similar to that of the Springer sandstones.

About 800 feet above the Jolliff is the Otterville limestone. In its southernmost exposures it is conglomeratic and similar to the coarse phases of the Jolliff, but over the remainder of the area it is a strongly bioclastic limestone.

TABLE 2

Jolliff Conglomerate Zone
measured on U. S. Highway No. 70, 3.2 miles west of Ardmore

		Top (feet)
1.	Shale, gray, fissile, with few one-half inch siderite layers	?
2.	Limestone, gray-brown, sandy to pebbly. Pebbles are black shale	2.5
3.	Shale, as above	7.0
4.	Limestone, sandy, with small shale pebbles	1.8
5.	Shale, as above	4.5
6.	Limestone, sandy, crinoidal, with prominent elluviation of limonite along bands perpendicular to the bedding	1.5
7.	Shale with thin beds of sandy limestone containing pebbles of shale and sandstone up to 1 inch in diameter	24.1
8.	Shale, gray to brownish gray, with sideritic partings	26.0
9.	Limestone, red-banded, sandy, hard and resistant	0.4
10.	Shale, gray	0.3
11.	Sandstone, red-banded, calcareous	0.7
12.	Shale, gray with common siderite partings	23.0
13.	Limestone, very coarse clastic with scattered small chert and shale pebbles up to 1 inch in diameter. Mostly crinoid debris	1.3
14.	Limestone, massive, coarsely crinoidal. Thin sandy zone in middle	4.0
15.	Shale, gray to brownish-gray, with scattered phosphate nodules	130.0
16.	Limestone, conglomeratic. Coarse clastic limestone made up largely of comminuted crinoid debris but containing pebbles of chert, limestone, sandstone, and shale. Modal size of pebbles less than 2 inches; maximum size, 10 inches	34.5
17.	Shale, gray, with siderite partings and phosphate nodules	7.0

At the base of the Lake Murray formation and approximately in the middle of Dornick Hills group is the Bostwick conglomerate, which, because it is a ridge former and because it is exposed in three belts, is the most prominent exposed unit in the Ardmore basin. It consists of limestone and chert conglomerate, sandstone, limestone, and shale with a maximum thickness of 500 feet and grading to a feather edge. Conglomerates are thickest and coarsest near the Criner Hills where cobbles up to six inches in diameter are common, and pebbles of all older Paleozoic formations now exposed in the Criner Hills (down to and including upper Arbuckle) are represented. In this area limestone pebbles and cobbles predominate over chert.

Northward from the Criner Hills sandstone becomes more prominent than conglomerate; the limestone pebbles disappear;

and the conglomerates become finer, thin, and eventually disappear at a point about five miles northwest of Ardmore. Southeastward the conglomerates also decrease in size and thickness, and limestone becomes progressively more prominent until it constitutes at least 75 per cent of the Bostwick beds in its southernmost exposures.

At its northeasternmost exposure, on the southeast nose of the Caddo anticline, the Bostwick consists of a bed of sandstone and thin limestone which is present only along two miles of strike.

The Bostwick truncates older Pennsylvanian beds on the east flank of the Criner Hills, but the truncation extends less than three miles from the Hills, and within the basin there is no evidence of an unconformity.

Above the Bostwick the Dornick Hills group consists of brownish gray shales with five or six limestone beds. Three of the limestone beds are named. These are the Lester, Frensley, and Pumpkin Creek members. The limestones are from about 20 to 50 feet thick, are predominantly coarsely bioclastic, and in many places are prominently cross-bedded. The strata from the top of the Frensley to the top of the Pumpkin Creek make up the Big Branch formation.

Between the Lester and Pumpkin Creek limestones at almost all places are two or three beds of sandstone. The sandstone is very light brownish-gray, fine-grained and massive.

One or more thin lenticular seams of coal are present below the Pumpkin Creek in the eastern part of the basin (sec. 17, T. 4 S., R. 4 E.) and have been mined for local fuel supply.

On the oil field structures of the northwestern part of the basin the Dornick Hills group is either missing or incomplete. Where the upper Dornick Hills is present, it onlaps over truncated Springer beds, as at Velma (Selk, 1951). In the structurally low areas more than 1,000 feet is present locally, but whether this represents the complete section is uncertain. The subsurface Dornick Hills consists of shales with interbedded limestones and conglomerates similar to the greater part of the surface section, but correlation of individual beds has not been generally possible.

Deese Group

Where exposed in the vicinity of Ardmore the Deese group consists of 5,300 to 8,800 feet of gray, gray-brown, and red shale in which are intercalated numerous massive sandstone beds, several thin limestones, and, locally, chert conglomerates.

Red shales are most common in the middle part of the formation, but thin red zones are found all through the Deese.

Sandstones comprise at least fifteen per cent of the Deese strata, and are slightly more abundant in the lower and middle portions than in the upper. The sandstones occur mostly in beds less than fifty feet thick, but thicker beds are present locally as in the thickest part of the Devils Kitchen member where nearly 300 feet of sandstone is present in two units separated by approximately 100 feet of shale. The sandstone is very light yellow-gray, fine-grained, friable, and well-sorted. Shale partings are rare, but cross-bedding is apparent on most weathered surfaces although it cannot usually be seen on fresh surfaces. Ripple marks, both symmetrical and asymmetrical, are common.

The beds are lenticular, and individual layers can seldom be traced more than a few miles. However, within the area of outcrop no regular increase or decrease in the aggregate amount of sandstone is apparent. About ten or twelve beds seem to be present in any given section, and they are markedly one very much like another.

Limestone occurs in numerous thin impure beds and in three beds which are thick enough and extensive enough to be mapped. The most prominent of these is the Arnold, which occurs near the middle of the Deese and can be traced over the entire outcrop area. It is at most places about 25 feet thick, but with interbedded shales reaches a maximum thickness of nearly 100 feet. It varies from nodular and earthy to coarsely bioclastic, and locally is a coarse crinoid coquina. It characteristically has lenses and nodules of smoky chert, some of which contain calcareous fossils.

About 900 feet below the top of the Deese is the Williams member, which consists of a few feet of silty limestone directly overlying approximately thirty feet of sandstone.

The Natsy limestone member occurs about 350 feet above the Williams. It consists at most places of four to six feet of limonite-stained fossiliferous limestone. Southeastward from Ardmore the

Natsy is partially conglomeratic, and north of Ardmore chert conglomerate occurs in the member correlated with the Natsy (Tomlinson, 1951).

Chert pebble conglomerates occur locally in at least eight different horizons within the Deese, but with the exception of the Devils Kitchen and Rocky Point members the conglomeratic units are all only fragmentarily preserved or exposed.

Over most of the area the Devils Kitchen unit is a massive sandstone carrying weathered chert grains which thins and wedges out to the northwest and northeast of Ardmore, but to the southeast thickens and develops into a coarse conglomerate. The conglomeratic phase becomes thicker and coarser as the outcrop swings to the east south of Lake Murray as far as exposures continue. At its easternmost outcrop in sec. 29, T. 6 S., R. 3 E., there is nearly 200 feet of conglomerate, sandy conglomerate, and pebbly sandstones in which chert pebbles two inches in diameter are common. The conglomeratic phase is exposed along about six miles of strike, and is present for an unknown distance to the east beneath the unconformably overlying Cretaceous sediments. Table 3 is a measured section of the conglomeratic phase of the Devils Kitchen.

The Rocky Point member is about 1,400 feet above the Arnold limestone. It is found only south and west of Ardmore, and a large part of its outcrop has been covered by Lake Murray. The Rocky Point has a maximum thickness of 70 feet, but much of the thickness is sandstone. Moreover, the sandstone becomes more dominant to the northwest, in which direction the unit thins and finally shales out in sec. 26, T. 4 S., R 1 E.

The conglomerates are almost confined to the area south of Ardmore; only two thin beds are known north of Ardmore and both occur at the approximate horizon of the Natsy limestone. The conglomerates increase in number, thickness, and coarseness to the south and particularly the southeast. As will be seen, this relation also holds for the overlying Hoxbar formation.

In the oil fields to the northwest of the outcrop area the Deese is much thinner than at the surface—commonly less than 2,000 feet. Just how much of the surface Deese is represented on these structures is not clear. In some areas Deese beds overlie lower Dornick Hills, Springer, or older beds and can be shown to onlap onto the

TABLE 3

Devils Kitchen Member
measured at Lake Murray spillway

		Top (feet)
1.	Chert conglomerate, fine-grained. Modal pebble size 5-6 mm, but range up to 40 mm. Matrix of fine- to medium-grained quartz sand. Bed is well-consolidated and makes a dip slope	2.0
2.	Sandstone and pebbly sandstone. Slightly brownish gray medium-grained very well-sorted sand, cross-bedded. Chert granules and small pebbles irregularly distributed in thin bands	6.0
3.	Chert conglomerate and pebbly sandstone. The conglomerate occurs in channels cutting into the sandstone. Pebbles are entirely chert, mostly light colored, but approximately 10 to 15 per cent dark gray or brown and 5 per cent light gray with dark laminae. Most of the pebbles are weathered white to a depth of 1-2 mm. Maximum pebble size 50 mm. Sandstone is fine- to medium-grained and very quartzose	13.5
4.	Silty shale, very light gray, with pebbles of chert	1.0
5.	Sandstone and pebbly sandstone. Sandstone is very light gray, fine- to medium-grained, very well-sorted, and cross-bedded. Pebbly zones occur in thin lenses in which the pebbles are rarely more than 4-6 mm in diameter. Chert grains are abundant in the sandstone	36.0
6.	Sandy conglomerate. Chert pebbles mostly less than 25 mm in diameter in a matrix of fine- to medium-grained quartzose sand. Dark cherts are much more abundant than in higher conglomerates. In some layers chert pebbles are conspicuously oriented parallel to apparent bedding	26.0
7.	Sandstone and slightly pebbly sandstone. Very light gray, fine- to medium-grained, very well-sorted, and cross-bedded	60.0
		144.5

uplift, indicating that the lower part of the section is missing. However, the evidence is not clear that this relationship holds where the upper part of the Dornick Hills is preserved. Furthermore, in the area just to the west of the Arbuckle Mountains the Deese is truncated completely by Hoxbar sediments, but the lateral extent of this unconformity is unknown. Some geologists correlate the subsurface Tussy limestone with the Arnold. The lithology of the Deese in the two areas is generally similar, even to the presence of the red shales. Thick chert conglomerates are present in the lower part of the Deese on the southwest flank of the Velma uplift, and thinner beds have been found in other areas.

Deese sandstones are important oil reservoirs in the area, with the Tussy and Fusulinid zones being probably the most important.

Hoxbar Group

The Hoxbar "group" comprises approximately 3,000 feet of strata similar to the Deese except that limestone is a more prominent constituent and sandstone less so. The basal member of the Hoxbar is the Confederate limestone, and the top is marked by the sharp angular unconformity at the base of the overlying nonmarine Pontotoc red beds.

The named limestone members are, in addition to the Confederate, the Crinerville, Anadarche, and Daube. These are spaced at nearly equal intervals of 400 to 500 feet. Each characteristically is the upper unit of a sequence of sandstone or conglomerate, shale, and limestone. Like the older limestones these are largely bioclastic. The Confederate and Crinerville carry chiefly a fusulinid and crinoid fauna; and the Anadarche and Daube, a brachiopod fauna.

Just below the Daube limestone in the northwest portion of T. 5 S., R. 2 E., a coal bed with a maximum thickness of four feet occurs. Several unsuccessful attempts have been made to mine this commercially.

The only named sandstone is the Zuckermann member, which lies about 500 feet above the Daube limestone. It includes up to fifty feet of sandstone occurring in four or five beds. For the most part it is a light gray fine-grained very limy sandstone, but at least locally it has at the base a bed of limestone and chert conglomerate.

In the part of the basin where it is known only from wells the Hoxbar is commonly about 1,000 feet thick, but reaches a maximum of nearly 3,000 feet. The amount of sandstone in the Hoxbar varies greatly from place to place, but individual members cannot be traced from uplift to uplift. However, the Hoxbar sandstones have been important oil reservoirs and include the prolific Healdton sand of the Healdton field.

Thin limestones occur at several horizons, and the lowest one has been correlated with the Confederate member of the surface section. In the area adjacent to the Carter-Stephens County line a limestone is present in the approximate middle of the Hoxbar which has a maximum thickness of 300 feet but thins laterally and is known over an area of only about 200 square miles. Characteristically it is thickest and most massive over the large anticlines which are the sites of the oil fields of this area. It has been named the County Line limestone. Wheeler (1947) has interpreted it as a reef deposit, and the petrography of the limestone seems to support this interpretation.

Pontotoc Group

The most abrupt change in the entire stratigraphic column is the contact between the steeply dipping marine Hoxbar and the nearly horizontal continental arkosic Pontotoc redbeds. The lowermost part of the Pontotoc group is characteristically a heterogeneous sequence of red arkose gravel and red silty shale. The gravel occurs in individual and coalescing channels, and the red shale presumably represents flood plain deposition. Most commonly the gravels are unconsolidated, but locally they are cemented by calcite. This sequence varies from a few feet to 100 feet thick, and is followed upward by red shale with lenses of red arkose and light gray quartzose sandstone similar to those in the lower beds. The Pennsylvanian-Permian boundary is conjectural as neither fossil nor diastrophic evidence of any break can be found. However, for convenience it is generally considered to be at the base of the unfossiliferous Hart limestone which is at the base of middle Pontotoc.

STRUCTURE

The Ardmore Basin is a long narrow downwarp trending northwest by west between the Arbuckle and Wichita Mountains. Its average width is about fifteen miles; and its length, at least fifty miles. To the northwest it continues with only minor interruption into the larger Anadarko Basin; to the southeast it disappears beneath a Cretaceous cover.

The general structure of the basin is synclinal; but a number of large anticlines occur within the basin; and the flanks are in considerable part faults of large displacement. Dips are very steep along the basin margins, and at Ardmore the outcropping strata are essentially isoclinal.

The greater part of the basin flanks is covered by a blanket of unconformable sediments, but from the large number of deep wells drilled in the area it is known that the south flank from the Criner Hills northwestward is a zone of high angle faults, and the north flank in the eastern part of the basin is similarly faulted. The north flank in the western part of the basin is not faulted, but is a belt of nearly vertically dipping beds. The general structure is somewhat like a propeller—asymmetrical with a faulted steep flank on the south in the western part of the basin and a faulted steep flank to the north in the eastern end.

Few of the individual faults have been delineated, but where the flanks are faulted it is known that the descent from the bordering highlands to the basin is in general by a series of steps, each step being made by a high angle fault with a displacement of several hundred to several thousand feet. The structural relief is great, and along the margins is apparently rarely less than 10,000 feet in a horizontal distance of three or four miles.

While the general movement of the faulting is downward toward the basin, there are local instances where an individual fault is downthrown toward the borderland. Particularly is this true in the southeastern part of the basin (in Marshall County). Here the margin is not well-defined, but rather consists of an area at least ten miles wide which is broken into a series of horsts and grabens.

The anticlines which occur within the basin are themselves large features, with several having a structural height of 5,000 feet or more. Those in the western part of the basin are the sites of several of the largest oil fields of Oklahoma; including, most notably, the Velma, Sholom Alechem, and Fox-Graham fields. The Velma and Fox-Graham fields (along with several minor pools) are located along a single major uplift known to be more than thirty miles long (Tomlinson, 1952). This uplift is a steep anticline which is locally faulted with a central horst (figure 13). Dips of sixty degrees or more are common on the flanks of the uplift, and in several places the beds are locally vertical. A part of the structural relief in the upper beds on the uplift is due to squeezing upward of the thick incompetent Caney and Springer shales on the flanks of the anticlines (Tomlinson, 1952).

In the outcrop area are two sharp folds, the Caddo and Overbrook anticlines, both with a structural height of at least 5,000 feet and with dips ranging up to vertical and locally overturned. In both of these anticlines the Caney and Springer shales have been squeezed upward over the core of the structure greatly increasing the tightness of folding (Tomlinson, *op. cit.*). Drilling has shown that in the Caddo anticlines the sharpness of folding decreases greatly with depth (Tomlinson, *op. cit.*), and it has been suggested (Westheimer, 1950) that the Overbrook anticline may actually be a rootless structure.

Sagging of the basin floor proceeded concomitantly with sedimentation, and in general the basin was filled as fast as it sank. Minor changes in depth of water are suggested by variation in fossil assemblanges from a dominant fusulinid fauna to a dominant brachiopod or gastropod fauna, and local emergence is shown by the two known occurrences of coal in the Dornick Hills and Hoxbar formations. However, the overall picture suggests extreme uniformity in the depth of accumulation—at least for the Dornick Hills, Deese, and Hoxbar groups. Except for the occurrence of coal locally at two horizons, these sediments seem to be all of shallow marine origin. Fossils are abundant throughout the section; the sandstones nearly all show the effects of strong current action either in cross-bedding or ripple marks; and the uniform light color of the sediments suggests a continuously aerated oxidizing environment.

During deposition of the Springer beds the Ardmore basin was part of a considerably larger sedimentary basin that covered all of southern Oklahoma and accumulated considerable thicknesses of detrital sediments (the Jackfork and Springer formations). There is no evidence that any of the uplifts now bordering the basin supplied any part of the Springer sediments, but it is possible that the basin was structurally differentiated from the surrounding areas at this time. The very extensive occurrence of slumped Springer sandstones on the flanks of the uplifts of the western part of the basin, particularly the Fox-Graham-Velma anticline, is interpreted to be the result of uplift during sedimentation. Whether the Wichita and Arbuckle Mountains were also tectonically active is not known. It seems probable that they were, but the evidence is equivocal.

The first definite evidence of sharp uplift of the basin margin and a local source of sediments is the Jolliff conglomerate in the lower part of the outcropping Golf Course formation. Similar Dornick Hills conglomerates are known from wells all along the south margin of the basin (Wheeler, 1949; Selk, 1951) and occur at several stratigraphic levels in the outcropping Dornick Hills beds.

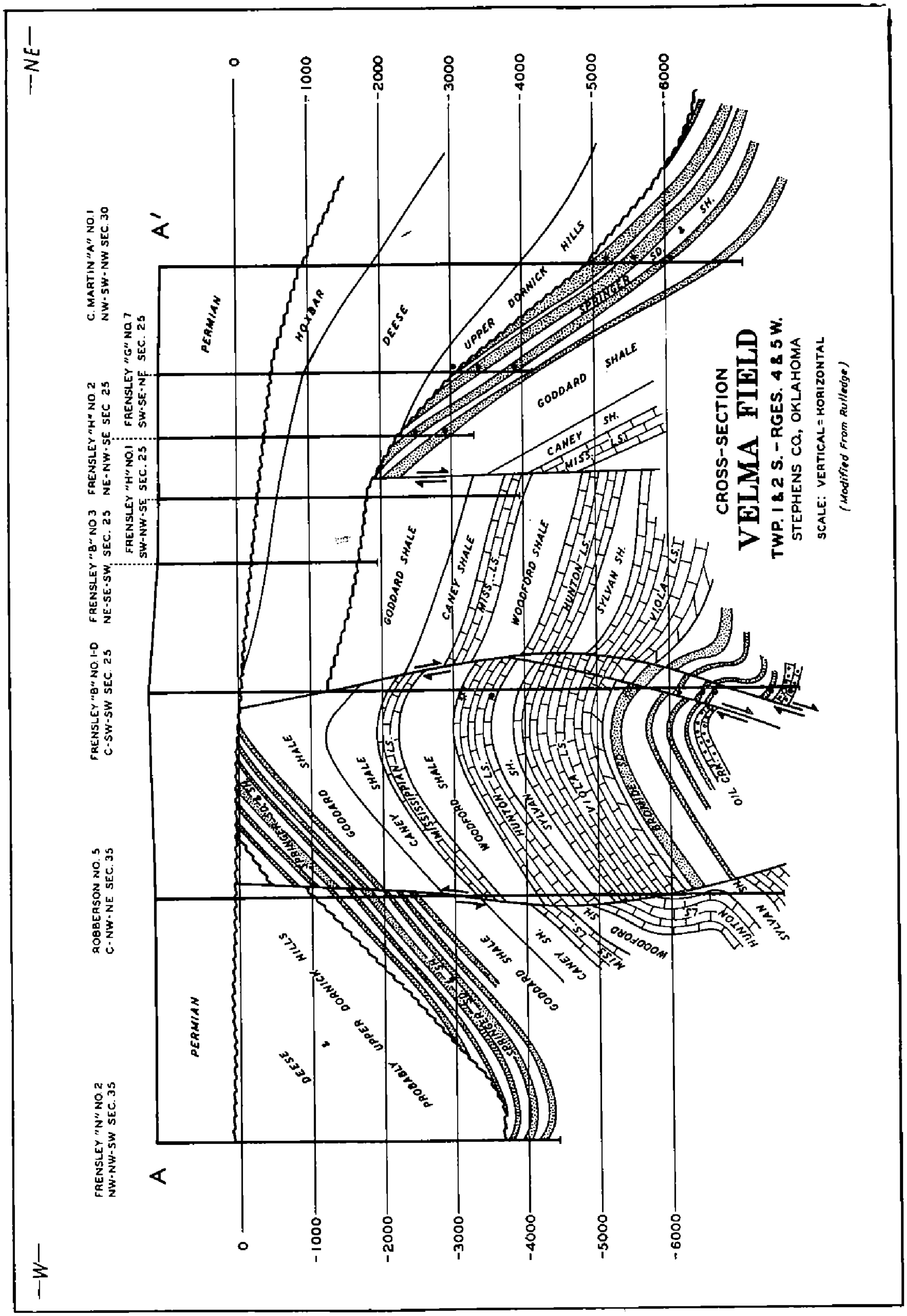


FIGURE 13. Structure section of the Velma Oil Field.

As there does not seem to be any continuous zone of conglomerate along the south margin of the basin, it is apparent that the basin did not subside as a whole, but as a number of segments each probably controlled by a single fault. The faulting was not concentrated in one or two short periods, but was spread throughout the time of sedimentation.

Along the south margin the more prominent conglomerates are in the Dornick Hills group though a few occur in the Deese. Along the north margin conglomerates occur at a number of levels throughout the Deese and Hoxbar formations; none is known from the Dornick Hills although it is possible this may reflect only the small area of exposure and the few wells that have been drilled in the area. However, the available evidence from the conglomerates indicates the subsidence of the basin was asymmetrical, that its early development was as a half-graben downfaulted on the south, and that the north margin was uplifted somewhat later.

The final deformation of the basin was brought about by moderate compression, probably resulting from final uplift of the Arbuckle Mountains which caused the squeezing of the Caney and Springer shales into the tight folds of the Caddo and Overbrook anticlines.

Unconformities occur at several levels in the stratigraphic section. There is no evidence of an unconformity between Mississippian and Pennsylvanian strata, but a widespread unconformity is present in the lower Pennsylvanian beds. This is evidenced by truncation of Springer and older beds on the flanks of the Wichita Mountains and on the anticlines of the western Ardmore Basin and surrounding area, and by the onlap of Big Branch, Deese, or younger beds onto the uplifts. As far as known, the unconformity is not continuous but is confined to the structurally high areas. Presumably the deeper areas have complete stratigraphic sections.

A second unconformity occurs at the base of the Hoxbar formation at the west end of the Arbuckle Mountains. In this area the Hoxbar truncates and onlaps all older formations from the Deese down to the Upper Ordovician beds. This unconformity is apparently local and reflects the gradual growth of the Arbuckle Mountains.

The most sharply angular unconformity of the Paleozoic strata is that between the Hoxbar and Pontotoc groups. The pre-Pontotoc beds are in many places around the Arbuckle and Wichita Mountains steeply folded and faulted while the Pontotoc beds are themselves flat or only slightly warped. This is true throughout the Ardmore Basin. Although the angular discordance is locally great, the interruption in sedimentation was minor; and it is probable the structural relief beneath the Pontotoc developed in part during pre-Pontotoc sedimentation, and not after that sedimentation came to a close.

PETROGRAPHY

Coarse Clastics

Distribution and Gross Features

Conglomeratic rocks are the most prominent stratigraphic units of the exposed Ardmore Basin. They are relatively resistant to erosion and characteristically crop out in strike ridges which may stand 100 feet or more above the surrounding countryside.

The more prominent outcropping conglomerates are the Jolliff and Bostwick members of the Dornick Hills group, and the Devils Kitchen and Rocky Point members of the Deese. Many lesser conglomerates are known from many wells drilled in the western part of the basin (Selk, 1951). Wheeler (1949) reported coarse conglomerate from the Dornick Hills near the Fox-Graham field, and at least two conglomeratic zones are present in the Deese formation of the Velma field. Probably the conglomeratic zones, where the pebbles are chiefly shale or limestone, are frequently not recognized in cuttings; and it is rarely that a core is taken.

The conglomerates occur interbedded with sandstone, shale, or even limestone, the whole constituting a relatively well-defined stratigraphic unit. In the thicker sections (Bostwick, locally 500 feet; Devils Kitchen, 300 feet) the conglomerate makes up less than half of the total (*see* Table 3). Individual beds of conglomerate range in thickness from thin stringers to massive beds up to 50 feet thick. The beds are lenticular, and have a maximum lateral extent of a few thousand feet. Obvious cut-and-fill structure is not common, but can be seen in several outcrops.

The conglomerates may have a considerable persistence in a direction parallel to the margins of the basin. For example, the Bostwick member can be traced for more than thirty miles. However, along this outcrop the thickness and coarseness of the conglomerate varies considerably.

In contrast to this persistence along the basin margins is the decided lack of persistence normal to the margins. The conglomerates are confined to a zone not more than ten miles wide along the edge of the basin; all thin, become finer-grained, and eventually disappear toward the axis of the basin.

Reconstruction of the shape and size of the conglomeratic units is limited by their fragmentary exposure and preservation. A linear strike outcrop exposes only a component of the three-dimensional stratigraphic relations and shape of a sedimentary body, much the same as an apparent dip offers only a component of the three dimensional structural relations. However, the Bostwick conglomerate, due to fortuitous stratigraphic position, crops out in three parallel strike ridges, each of which exposes a cross-section of the unit.

The Bostwick conglomerate is a wedge-shaped basin margin deposit. The conglomerate is thickest (up to 500 feet) and coarsest near the south margin, and does not change greatly parallel to the margin. In the successive outcrops toward the center of the basin the conglomeratic zone rapidly becomes thinner and changes character. Limestone and chert conglomerates give way to chert and sandy chert conglomerates, and these in turn to cherty sandstones. The outer edge is marked by a discontinuous zone of sandstone.

The Devils Kitchen is on the opposite margin of the basin from the Bostwick. Its main outcrop is a long curving band well out in the basin. For most of its length it parallels the basin margin and exposes only pebbly sandstone; but to the southeast of Ardmore the outcrop curves to the east, and the conglomerates appear and rapidly thicken and become coarser as far as the formation is exposed.

Because the thickest part of the wedge is not visible and because most of the outcrop is near the center of the basin, the Devils Kitchen is not as prominent as the Bostwick member. Nevertheless, it is almost certainly a considerably larger sedimentary unit.

The other conglomerates which are exposed are only small fragments of the original deposits. Probably they are all remnants of similar marginal accumulations; but if so, the upper parts of the wedges have been eroded. It is also probable that still other conglomerates were present, but have been completely destroyed by erosion of the margins of the basin.

Texture

The conglomerates are not notably coarse, and the greater part of them are better termed sandy conglomerates or even pebbly sandstones. All become coarser toward the basin margins; and as none is exposed completely to the margin, presumably the coarsest conglomerates have been eroded or are buried beneath younger sediments. The largest fragment seen was a one-foot shale boulder in the Jolliff member; limestone cobbles 6 inches in diameter are present in the Jolliff and Bostwick outcrops nearest the Criner Hills; and Schacht (1947) has reported a 7-inch chert cobble from the Devils Kitchen. These are, however, rare occurrences; and the modal pebble size is probably not more than ½ inch.

Except in one outcrop of the Jolliff and one of the Bostwick, both of which are very near the perimeter of the basin, all of the conglomerates have a sandy matrix. The proportion of matrix increases in a basinward direction in all occurrences, and the conglomerates gradually give way to pebbly sandstone. Generally, the pebble size decreases as the amount of sand increases.

The pebbles and sand constitute two sharply differentiated size distributions; the two overlap hardly at all. As a result, the characteristic size distribution of the conglomerates is strongly bimodal. Examples are shown in figure 14.

All ratios of pebbles and matrix may be found, and in many of the occurrences the pebbles are not in contact with one another but are scattered as in the proverbial "plums in a pudding".

In the coarser conglomerates the pebbles are very poorly sorted, but in the medium and fine conglomerates the pebbles are moderately well-sorted (figure 14).

With few exceptions the pebbles and cobbles are sub-angular to subrounded. Mean roundness of fifty cobbles (10 to 25 mm) from the Devils Kitchen member, as determined by visual comparison (Krumbein, 1941) is 0.51; and mean roundness of the same number of cobbles from the Bostwick is 0.58. Observed differences in roundness are thought to be chiefly due to the differences in the source of the pebbles, rather than differences in the abrasion history. A large proportion of the Devils Kitchen pebbles show a derivation from a source rock with a blocky fracture. These pebbles are irregular polyhedrons with only the corners rounded. On the other hand, many pebbles in the Bostwick were derived from nodular chert in the lower Paleozoic limestones and inherited a considerable degree of primary roundness.

The observed roundness of the pebbles is in keeping with the interpretation of the conglomerates as basin-margin accumulations. Krumbein (1940, 1942) has shown that rounding of angular fragments to the degree found in the Ardmore conglomerates can be accomplished by only a few miles of transport. Although the rate of rounding decreases as roundness increases, the subangular pebbles of the Devils Kitchen suggest that the distance of transport was short.

No attempt was made to measure orientation of the pebbles in the conglomerate, and in general there is no obviously visible preferred orientation. However, in a few occurrences elongate pebbles are highly oriented. An example of such a conglomerate from the Bostwick member is shown in figure 15.

Bonding of the conglomerates is by a chemical cement which may be calcite, secondary quartz, or limonite. Calcite is the most effective and the only one which is quantitatively significant. It is well developed in all of the conglomerates which have limestone pebbles, and is absent in those without limestone pebbles.

Quartz overgrowths are present in most of the non-limestone conglomerates, but they are small and barely sufficient to make the rock cohesive.

Limonite is a minor cement; and where it does occur, is concentrated in irregular zones. The limonite cement may form irregular masses; or where it is concentrated along intersecting planes, may form a false box-work structure. Both occurrences are fairly common, but particularly good examples are to be seen in the Bostwick northwest of Ardmore.

The limonite is secondary, and is probably a product of post-diagenetic weathering or groundwater circulation. The source of the limonite is most likely the iron-bearing carbonate and impure chert pebbles within the conglomerate.

Mineral Composition

All of the pebbles in the conglomerates are sedimentary rock fragments. Included are chert, limestone, shale, and sandstone; but chert and limestone make up well over 99 per cent of the pebbles in the outcropping conglomerates.

In the exposed part of the basin shale pebbles are most common in the Jolliff member, but even here they do not comprise more than one per cent of the pebbles. However, in one conglomerate core from the Velma area shale was the dominant pebble type. The shale which makes up the pebbles is very dark gray, smooth, and unctuous. In several of the larger pebbles brown siderite bands up to one-half inch wide are present. This shale of the pebbles is similar to that of the Springer formation, and is probably detritus having its source in the Springer.

Sandstone pebbles are even rarer than shale pebbles, but they can be found in all of the conglomerates. The largest one seen was in the Devils Kitchen member and had a greatest diameter of 1½ inches; it was a thin-bedded fine-grained angular quartzose sandstone. Smaller pebbles of very similar sandstone are rare constituents of the Jolliff. In the Hoxbar conglomerates there are small pebbles of quartz-cemented very well-rounded sand grains.

Limestone is a major pebble type in the conglomerates exposed near the Criner Hills, and all of the formations now cropping out there are represented in the pebbles of the Bostwick and Jolliff conglomerates (Tomlinson, 1929). In the Jolliff, pebbles of the Sycamore and Hunton limestones predominate; and in the Bostwick, pebbles of Viola, Simpson, and Arbuckle limestones are most common.

The proportion of limestone pebbles decreases rapidly away from the source, and they are largely confined to a zone parallel to the basin margin and not more than three or four miles wide. Limestone pebbles are found only in rocks which are thoroughly cemented.

In most of the conglomerates chert is either the dominant or the only type of pebble. The proportion of chert increases as the detrital limestone decreases, and in the basinward occurrences all of the pebbles are chert.

Most of the chert pebbles can be classed as either light or dark colored, and there is little overlapping between the two types. The light colored chert is very pale gray or yellowish gray, and much of it is finely laminated. The dark chert is nearly all dark brownish gray, and commonly is faintly mottled. The two colors of chert occur together, but in any single occurrence one is very likely to predominate strongly over the other.

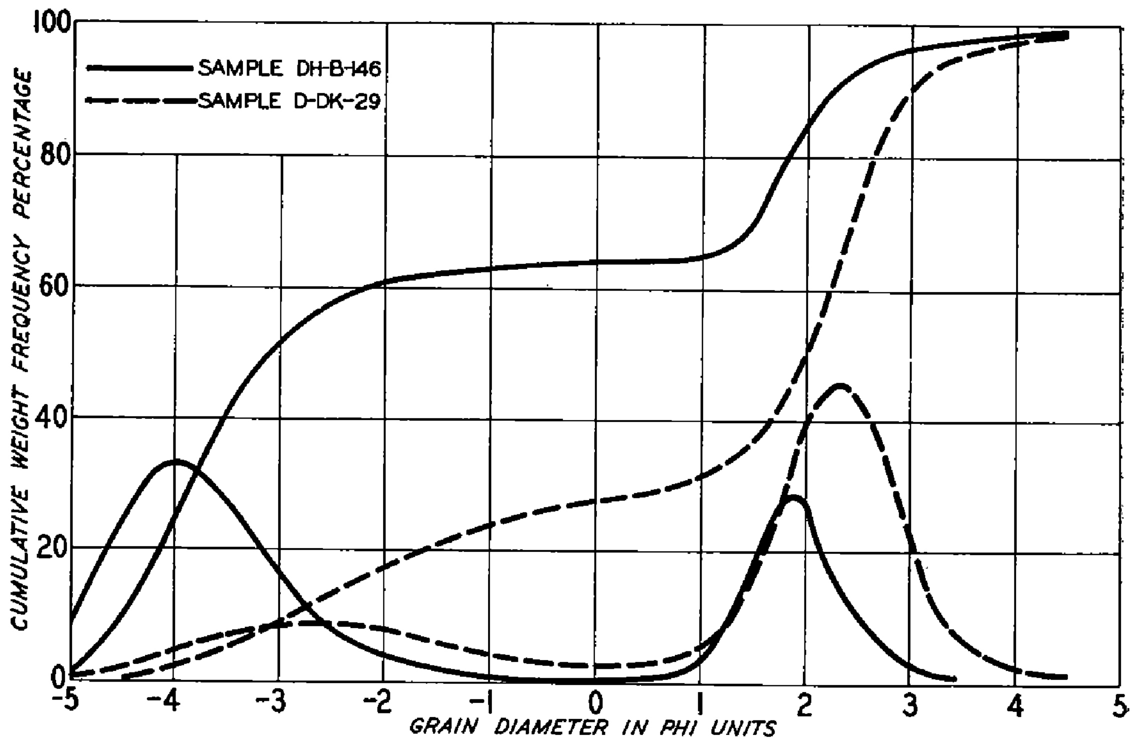


FIGURE 14. Unique frequency and cumulative frequency curves for two conglomerate samples.

A few pebbles of chert breccia were found in the Bostwick conglomerate.

In a number of occurrences, and particularly in the Devils Kitchen conglomerate, weathering of the chert has given the pebbles a rim several millimeters thick of white tripolized chert (figure 16). This is almost certainly pre-depositional (or possibly intrastratal) weathering as it is not confined to the surface exposed at the present time.

As shown in thin section the most distinctive type of chert is made up of a mass of sponge spicules (figure 17). Most, if not all, of the laminated chert seems to be of this type.

Another microscopic variety consists of very fine but uneven grained microquartz. A small amount of opaque dust is scattered throughout, and irregularly shaped holes are very common, possibly the result of leaching of carbonate minerals. Irregular calcite or dolomite inclusions were seen in a few pebbles from the Bostwick conglomerate.

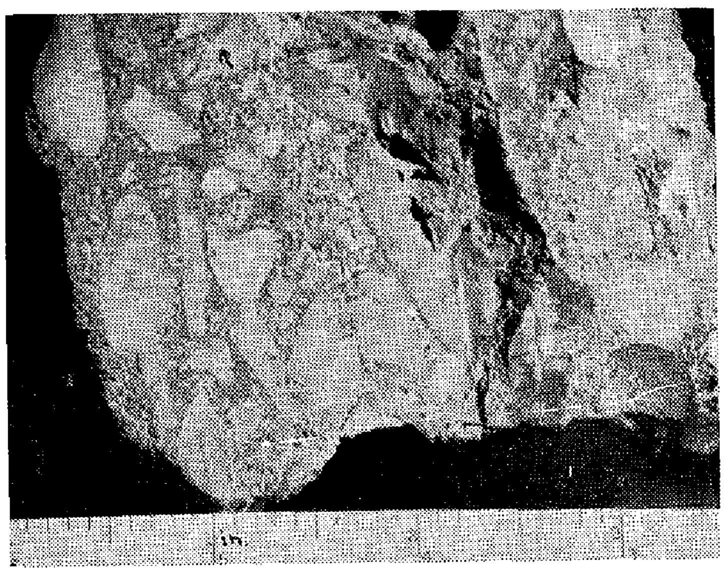


FIGURE 15. Bostwick conglomerate, hand specimen.

The dark colored chert is slightly clayey, and in transmitted light the fresh chert is colored in some shade of yellow-brown; where weathered, it is nearly opaque and isotropic. A few small round bodies which may be radiolarian skeletons are present, and rarely a silt-sized quartz grain can be seen.

Differences in the type of chert between the several conglomerates are not great. The most apparent difference is the degree of weathering; those conglomerates on the eastern margin of the basin have chert which is appreciably more weathered than those on the west. There seems to be slightly more dark colored chert in the Devils Kitchen conglomerate than in the Bostwick, but the difference is not great. All chert types found in the Devils Kitchen conglomerate can also be found in the Bostwick.

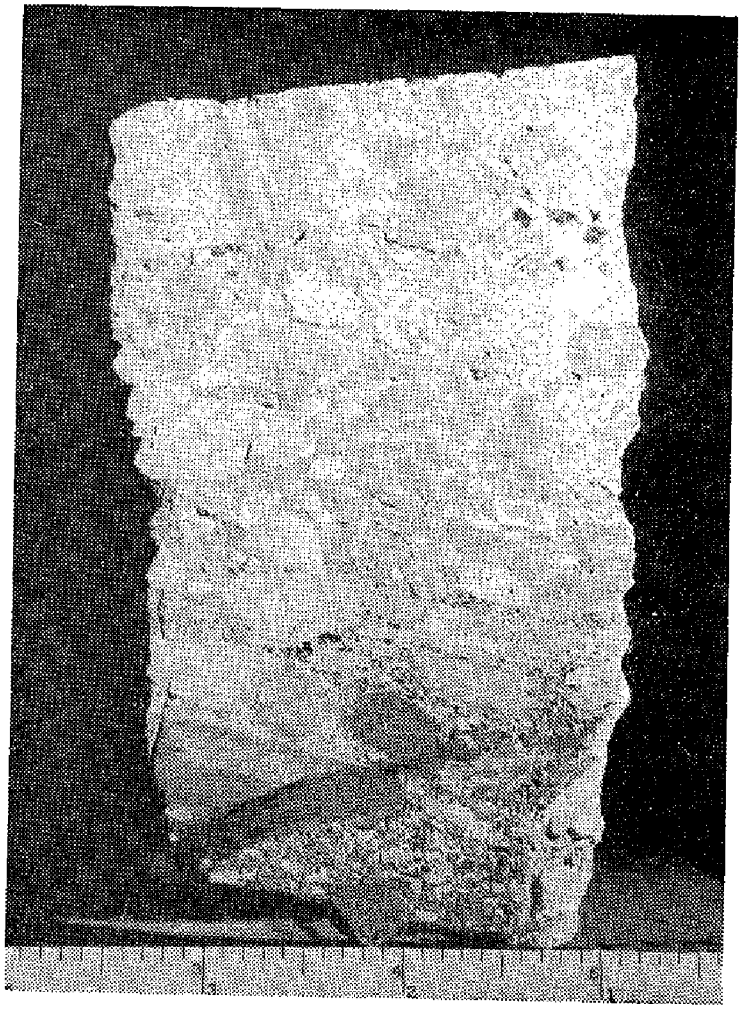


FIGURE 16. Devils Kitchen conglomerate, hand specimen.

FIGURE 17. Spiculitic chert pebble from Bostwick conglomerate (plane polarized light).

SANDSTONES

Gross Features

Bedding.—The sandstone zones of the Springer formation range from thin rhythmic interbedding of sandstone and shale to thick massive layers of sandstone. Cross-bedding occurs on a microscopic scale, but no large scale cross-bedding was seen.

The most common occurrence of the Springer sandstones is an alternation of thin beds of sandstone and shale. A typical occurrence is shown in figure 18. The beds are not gradational with one another, and the bedding is rhythmic rather than graded.

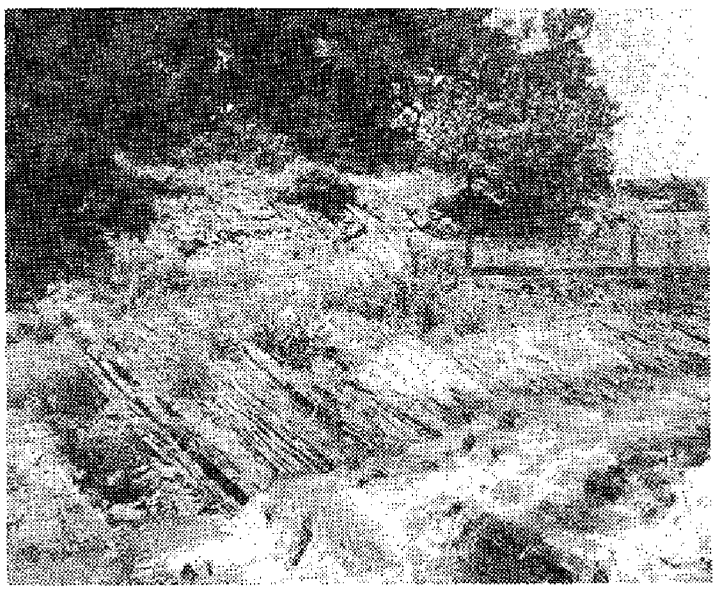


FIGURE 18. Outcrop of Overbrook sandstone at Lake Ardmore spillway.

Massive beds occur in two rock types, nearly pure quartz sandstone and sandstone with a clay matrix. Both of these types are gradational with the more common thin-bedded sandstones, and the three together are interpreted to represent parts of a single frequency distribution of bedding. The massive sandstones with the clay matrix are believed to be the result of accumulation without any essential modification by current action; the thin-bedded sandstones have resulted from moderate current action, and with increasing winnowing the shale interbeds become less prominent, and the sandstones cleaner and more massive. The most massive non-clayey beds are, thus, the product of the most intense and long continued current action.

The massive winnowed sandstones occur in the outcrop area along the flank of the Arbuckle Mountains and spottily along the

Caddo and Overbrook anticlines. In the buried part of the basin, the important occurrences of these sandstones are on the crests and flanks of the anticlinal areas where they form the very productive main reservoirs of the Velma, Sholom Alechem, Fox-Graham, and lesser oil pools. On these anticlines the clean massive sandstones grade into the thin-bedded type on the flanks, and the edge of the reservoir is determined at least in part, by the diminished porosity of the sandstones as their clay content increases (Davis, 1951).

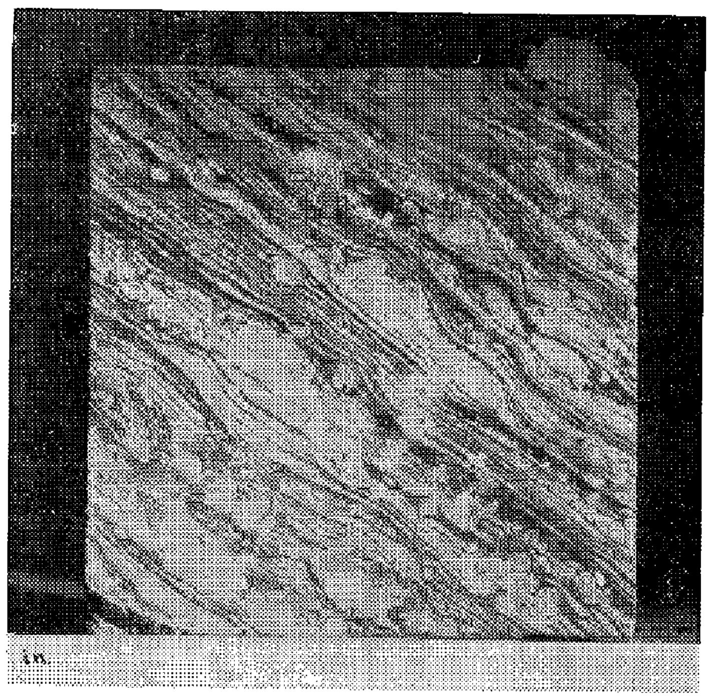


FIGURE 19. Disturbed bedding in Springer sandstone.

Slumped bedding is common in the Springer sandstones, and several examples are illustrated in figures 19 and 20. A number of types of disturbance can be recognized. The most common is slight crinkling of the bedding, and this is almost universal in the sandstones with thin clay laminae. The individual beds are thrown into folds with amplitude of a few millimeters to about one centimeter. The tightness of folding varies greatly. In most occurrences the folding is parallel through several beds, but it may be confined to a single layer. The folds are not eroded on the anticlinal crests, which suggests the slumping took place below the surface of sedimentation, and some time after the sediment involved in the slumping was deposited.

A second type of slumping is more nearly analogous to faulting than folding, as movement has occurred along sharply defined slippage planes. This is much larger scale disturbance and has resulted in the displacement of large masses of rock, and brought into juxtaposition widely different rock types. A good example is shown in figure 20. This sort of disturbance was not seen anywhere in the outcrop area, but it is very common on the flanks of the steeply folded anticlines of the western part of the basin.

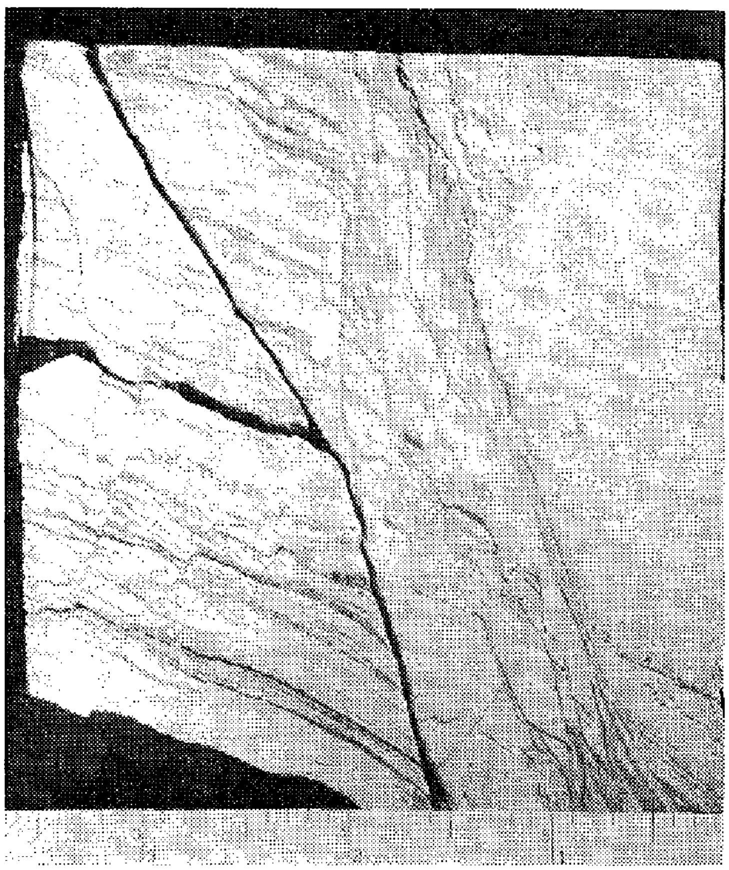


FIGURE 20. Slumped bedding in Springer sandstone. Note difference in texture and composition between far right side and remainder of core.

Still another type of slumping is complete disruption of the bedding due to flowage. This, also, is confined to the anticlinal flanks of the western Ardmore Basin. In some instances the bedding has been broken and a small or large mass rolled into a spiral; in other instances the orientation is chaotic and the rock consists of rounded masses of sandstone, usually less than a centimeter in diameter, irregularly distributed through a clay matrix. It is striking that in all occurrences of slumping by flowage the sandstone and shale have not mixed; it seems probable that at the time of slumping the two had such radically different physical states that they were unable to mix.

These three types of slumping, though distinctive, are only end members. To a limited extent they are gradational with one another, and quite commonly two types occur together.

All forms of slumping are apparently related to the occurrence of "clay" in the sandstones. Insofar as is known, slumping in the Springer sandstones occurs only where the proportion of clay matrix is high (about 10 per cent or more) or where there are closely spaced clay laminae. It seems apparent the essential slippage was in the clay or that the clay served as the lubricant along which the sand masses moved. In the interbedded sand and clay the sandy layers have only broken, they have not themselves flowed. It is only in the rocks with a clay matrix that movement between individual grains is apparent.

It is probable this large scale slumping has caused some of the erratic variations in thickness noted in some wells drilled on the anticlines; and it is possible that slumping has contributed to the thinning of the Springer formation over the crests of the anticlines though this is usually attributed entirely to post-Springer pre-Deese erosion.

The common occurrence of slumping in the Springer sandstones suggests the presence of local slopes within the basin of deposition, and also is evidence that the sediment remained in plastic or potentially plastic state for an appreciable time after deposition. Because all occurrences of the most intense slumping are known only from core samples the size of the slumped masses cannot be determined. However, in one well, the Gulf No. 3 Pruitt in the Fox-Graham area, cores and the electric log suggest at least a 200-foot thick section of rock has slumped. The Springer clays are very fine-grained and contain appreciable amounts of montmorillonite. It is likely these clays were sedimented as thick colloidal suspensions, probably with thixotropic properties. Thus, the clays may have existed in a gel state for long periods, only suddenly to become fluid when disturbed by uplift of the anticlines (Boswell, 1949).

In the sandstones of the Dornick Hills, Deese, and Hoxbar groups bedding is simple and nearly uniform throughout the stratigraphic section. The sandstone occurs in layers a few feet to fifty feet thick which have only very minor bands of shale or are, as far as can be seen, completely unbroken. Superficially, these thick beds appear massive, but most of them can be seen to be obscurely cross-bedded on a large scale. The cross-bedding occurs in subparallel beds from about a foot to as much as ten feet thick, and the laminae are inclined at an angle of about fifteen degrees to the base of the bed. In any single exposure there is general uniformity of the direction of inclination, but over the basin the variation is great enough to make obscure any pattern which may be present.

Insofar as is known, slumping has not occurred in these younger sandstones.

All of the sandstone beds in all of the formations are strongly lenticular, and the scale of lenticularity varies from microscopic in small scour and fill structure to thick massive sandstone bodies which extend for as much as 10 or 15 miles. The individual lenses seem all to be tabular (Krynine, 1948), and the ratio of width to thickness is of the order of several hundred to one. In the Springer formation particularly, and to a lesser extent in the other formations, the sand bodies occur in relatively well-defined sandy zones with a lateral extent much greater than the individual lenses. While the individual sands cannot be followed for more than a few miles, the sandy zones of the Springer are known from close well control to be continuous over an area of at least 300 square miles and possibly may be present over a considerably greater area.

MINERAL COMPOSITION
(major constituents)

Technique.—Study of the major mineral constituents was mostly from thin sections; these were supplemented by X-ray determination of the clay minerals present. An attempt was made to determine the quartz-feldspar ratio from study of loose grains with the staining technique described by Russell (1935). However, the amount of feldspar was less than the variation due to experimental errors; so the results have not been used.

One hundred and ten thin sections were examined; for all of these the constituents were identified and described, and some estimate made of their frequency. For thirty thin sections mineral frequencies were determined by the point-count method as described by Chayes (1949) and are presented in Table 4. Four hundred points were counted for each sample.

As variation of a percentage determination has a binomial frequency distribution (Snedecor, 1946), the precision of the percentage estimation varies not only with the total number of grains counted but also with the abundance of the mineral. The more abundant minerals are determined with greater precision than the less abundant. Figure 21 shows the 95 per cent confidence interval for any percentage based on a count of 400 grains. This graph is applicable to the percentages given in Table 4. For example, in sample D-112 for which the quartz content was estimated as 78 per cent, the probability is nineteen to one that if the count were continued indefinitely, the quartz percentage would be somewhere between 74 and 82.

For purposes of counting, the mineral constituents were divided into eight groups. Quartz, feldspar, and chert were the only individual minerals differentiated. The rock fragments were divided into shale and/or slate and other rock fragments except limestone and dolomite. All carbonate was considered as one group regardless of mineralogy or origin. The heading "matrix" is a textural rather than compositional class as it includes fine-grained interstitial mineral matter; for the most part it consists of clay minerals, but it also includes other minerals not resolvable with the microscope. The miscellaneous column includes the micas, glauconite, barite, collophane, kaolinite (as mud balls), gypsum, and the "heavy" minerals.

Description of results:

The Ardmore basin sandstones are composed mostly of stable constituents. They are relatively quartzose, and their feldspar and mica content is extremely low. No definitely identifiable igneous rock fragments and very few metamorphic rock fragments were found, and it is clear the major part of the detritus is of direct sedimentary origin.

Mineral composition of thirty samples selected to show a full range of rock types is shown in Table 4.

TABLE 4

MINERAL COMPOSITION OF ARDMORE SANDSTONES

Sample No.	Quartz	Shale frags.	Other rock frags.	Car-bonate	Feld-spar	Chert	Matrix	Misc.
			Springer outcrop					
S-RC-45	85.4	0.4	0.6		0.2	0.2	12.6	0.6
S-RC-175	52.6			43.4		1.2	1.4	
S-O-178	98.8				0.2	0.2	0.8	
G-198	87.4	0.6	0.8		0.2	0.4	7.4	3.2
			Springer cores					
S-S-77	83.2	2.8	0.2	3.8	1.0	1.8	5.0	2.2
S-H-80	86.6		1.0	0.6	0.4		11.0	0.4
S-G-402	89.2	1.0		1.0	0.4	0.4	6.6	1.4
S-403	80.4	0.4		17.2	0.8	0.6	0.4	0.2
S-417	85.2		0.6		0.4		7.4	6.4
S-423	49.0		1.0	45.2	0.6	0.4		3.8
S-S-457	90.4	1.6	0.8		1.4	0.2	3.0	2.6
S-S-464	64.0				0.4	0.4	32.0	3.2
			Dornick Hills outcrop					
DH-36	81.6	6.8	5.2		2.2	0.4	3.8	
DH-B-43	99.4					0.6		
DH-140	61.4		1.6	35.0	1.0	1.0		
DH-186	77.2	7.4	8.8		2.4	0.8	3.2	0.2
DH-B-240	55.2			36.2	3.4	5.2		
			Deese outcrop					
D-DK-25	92.2	0.6	2.6		0.8	1.2	2.2	0.4
D-32	79.2	9.4	6.4		1.4	0.8	2.0	0.8
D-DK-40	82.8	1.6	1.0		2.0	4.4	8.0	0.2
D-N-98	79.7	8.0	5.3		3.5	0.5	0.8	2.1
D-112	78.0	10.6	3.8		2.8	0.2	4.4	0.2
D-176	89.2	4.5	2.2		1.1	0.4	2.2	0.4
D-234	56.6	20.0	12.0		1.6	1.2	5.0	3.6
			Deese cores					
D-284	81.2	7.4	4.4		4.2	0.2	2.2	0.4
D-285	70.4	10.4	3.0	7.6	2.4	0.8	4.0	1.4
			Hoxbar outcrop					
H-34	82.2	7.0	4.8		0.8	0.4	4.6	0.2
H-84	59.8		0.4	33.2	0.8	2.0		0.6
H-Z-90	69.2			30.8				

The quartz content of the sandstones ranges from 49 per cent to 99 per cent; and the only other major constituents are rock fragments of various sorts, carbonate minerals, and clay minerals. The rock fragments are the most distinctive feature of the sandstones. They are unusual in that they consist for the most part of rocks which are very susceptible to destruction during weathering and erosion.

Rock fragments to be expected in multi-cycle sediments such as these are such resistant types as metaquartzite and quartzose schist. Instead, in the Ardmore sandstones the common rock fragments are shale, limestone, and clayey siltstone. These rock fragments are not likely to have passed through several cycles of sedimentation; and, hence, very probably were added to the detritus in the present cycle.

The sandstones all have in common a low feldspar, mica, and chert content; they vary chiefly in the amount of rock fragments, carbonate, and matrix present. Several distinctive types can be differentiated, and to a limited extent these follow stratigraphic boundaries.

The Springer sandstones are on the whole the most quartzose, and a large part of them are orthoquartzites. They have a very low content of detrital rock fragments; the only major silicate detritus other than quartz is clay, and there is a complete gradation from the nearly pure quartz sandstone (samples S-178 and S-437) to very shaly sandstones (sample S-464). Carbonate is an abundant constituent in several samples, and it occurs either as authigenic chemical cement, either calcite or siderite, or as bioclastic grains of local derivation.

The Dornick Hills, Deese, and part of the Hoxbar sandstones are generally characterized by their relatively high content of rock fragments (up to 32 per cent). A few of the sandstones (DH-B-43, D-DK-25) have a low content of rock fragments and are similar to the quartzose sandstones of the Springer; but as these exceptions are found only in the conglomeratic zones, it is suspected that the rock fragments have been destroyed by the intense current action associated with the transportation and deposition of the conglomerates.

The greater part of the Hoxbar sandstones from the outcrop area form a distinctive rock type (H-84, H-Z-90), which is distinguished by its high proportion of carbonate and its low content of shale and other rock fragments. The carbonate of these rocks occurs as grains (figure 22), and is clearly of detrital origin as it includes a number of limestone types. The quartz in these sandstones is also distinctive in that it consists of very well-rounded grains in contrast to the angular fragments of the older sandstones.

On the basis of their detrital silicate mineral composition the Ardmore sandstones can be classified as orthoquartzites, protoquartzites, and quartzose graywackes (with possibly a few normal graywackes). The quartzose graywackes occur in two distinctive classes: (1) relatively coarse-grained sandy rocks rich in rock fragments, and (2) relatively fine-grained silty rocks rich in matrix ("clayey sandstones").

Protoquartzites and orthoquartzites.—Sandstones in which the only major detrital silicate is quartz are found in all formations. Examples in Table 4 are S-437, DH-B-43, D-DK-25, and H-Z-90. In the Springer formation quartzose sandstones occur in all members as isolated lenses which grade laterally into less quartzose types. In the Dornick Hills and Deese formations highly quartzose sandstones were found only in the conglomeratic members. Here they occur interbedded with the conglomerates, and they make up the basinward fine-grained borders of the conglomeratic wedges. In the Hoxbar formation most of the sandstones of the outcrop area are orthoquartzites, but those in the buried western part of the basin are not. Owing to the sampling problem, the proportion of these highly quartzose sandstones is impossible to determine with any accuracy, but a rough estimate would be that 15 to 20 per cent of sandstones of the basin have 90 per cent or more of quartz in their detrital silicate fraction. The amount of such rocks in the Dornick Hills and Deese formations is insignificant, but they make up probably at least half of the Hoxbar sandstones, and possibly close to a quarter of the Springer sandstones.

The quartz which makes up the bulk of these rocks is generally similar throughout the Springer, Dornick Hills, and Deese formations; but is sharply different in the Hoxbar orthoquartzites. The two assemblages of quartz differ both in shape and internal morphology.

The Springer-Dornick Hills-Deese quartz occurs as angular to sub-angular fragments, all with a pock-marked or pitted surface (figure 23). All grains contain abundant inclusions, by far the most common of which are irregular vacuoles and minute shreds and dust-like particles too small to be identified. Inclusions may be so abundant as to give a clouded appearance to the grain, or be distributed in irregular patches. In about a third of the grains the cavities show some degree of lineation, mostly as short slightly curved subparallel lines. Microlites of identifiable size are relatively rare, occurring in not more than ten per cent of the grains. Apatite is the most common; but also found are zircon, tourmaline, biotite, muscovite, orthoclase, titanite, and needle-like inclusons which are probably rutile.

The majority of the grains are fragments of single crystals which show a normal to very slightly strained pattern of optical extinction. However, an appreciable number are compound grains with an interlocking granular texture. These grains are probably fragments of vein quartz (Adams, 1920). The constituent crystals have different optical orientations, but they are markedly clear of strain shadows, and the internal boundaries are sharp and smooth. Rows of dust-like inclusions are common along the contacts, and a few grains have large inclusions of orthoclase and muscovite. In many grains all of the individual crystals are approximately the same size, but some show great variation with a non-seriate size distribution. In these latter grains fine quartz fills the interstitial spaces among the coarse crystals. The shape of the individual crystals ranges from equant to strongly elongate, possibly due to differing orientations of the grains. If they are elongate, their orientation is subparallel; and they form a characteristic "comb" structure.

This vein quartz, as would be expected, has an appreciably coarser size distribution than do the single crystal grains; and, hence, is most prominent in the coarser-grained sandstones. In the average sandstone not more than 10 per cent of the grains are of vein quartz; but in the coarser samples, such as S-78, S-423, and D-DK-25, 35 to 45 per cent of the grains are vein quartz.

The quartz of the Hoxbar quartzites contrasts in many respects with that of Springer-Dornick Hills-Deese sandstones. The grains are exceedingly well rounded and have a smooth or frosted surface (figure 24). Visible inclusions are present in nearly every grain, but they are few and scattered and never reach the abundance characteristic of the other quartz. Further, there is none of the vein quartz which is prominent in the other quartzites (except that some of the Hoxbar sandstones have a small admixture of the angular quartz). In its extreme modification this well-rounded Hoxbar quartz is characteristic of true polar orthoquartzites, either first or second cycle. It is similar to the quartz of Simpson sandstones (Ordovician) of the Arbuckle Mountains (figure 9).

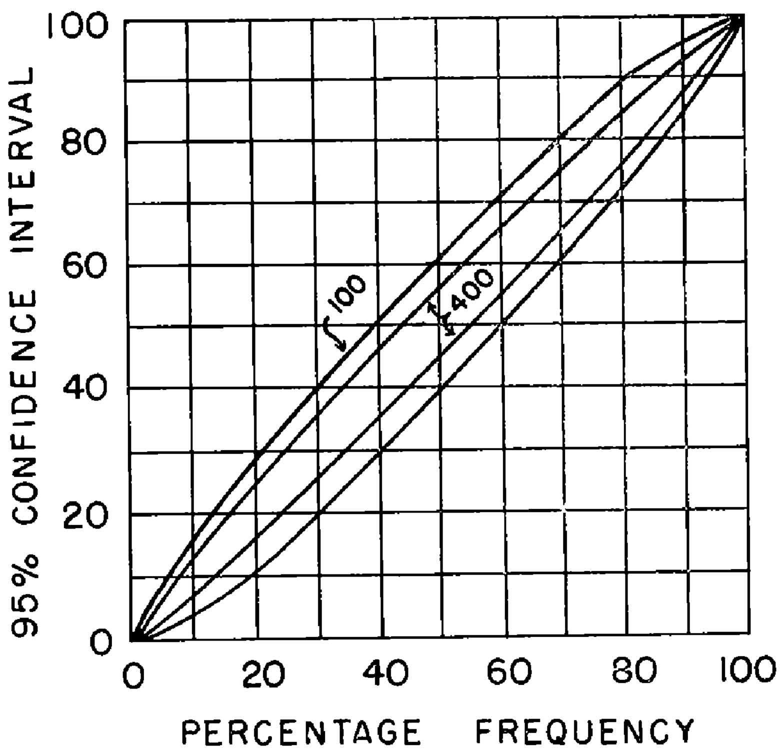


FIGURE 21. Ninety-five percent confidence interval for binomial distribution (n = 100, 400).

Quartz overgrowths occur in almost every sample of the orthoquartzites, but they are usually small and do not effectively bind the rock. The largest overgrowths occur on scattered grains in the Hoxbar sandstones; however, these are broken and worn, and are likely inherited from a previous cycle of sedimentation.

Other than quartz, carbonate minerals are the only major constituents of the orthoquartzites; and they occur mostly in the Springer and Hoxbar formations and only in a few instances in the Dornick Hills formation. Included are calcite, dolomite, and siderite. There are several modes of occurrence of the carbonate: (1) as detrital grains which have been transported from outside the basin, (2) as clastic grains, mostly fossil fragments but including some oolites, which have originated within the basin, and (3) as a chemical content.

The detrital carbonate is confined to the Hoxbar formation, but it may constitute up to 25 per cent of individual sandstones as in samples H-84 and H-Z-90 of Table 4. The carbonate occurs as limestone and dolomite fragments of a wide variety of types. Included are grains of the distinctive silty dolomitic limestone of the Sycamore formations, and other grains are similar to the various limestone and dolomite formations of the Arbuckle mountains. The detrital nature of the grains is shown by the much greater degree of weathering of the grains than of the calcite cement which usually is present and by the wide variety of rock types represented. In most samples the calcite has been partially redistributed and crystallized to a sparry cement, but relict boundaries of the grains are usually preserved. In a few occurrences the calcite has not recrystallized, and the sandstone is completely friable.

The few limy sandstones of the Dornick Hills formations (DH-140 and DH-B-240 of Table 4) may also contain detrital limestone. At least the presence of relict grains boundaries suggests a clastic depositional texture, but recrystallization has gone so far as to make determination of the origin of the grains impossible.

The clastic fossil fragments and oolites were found only in the sandstones of the Springer formation; and as far as is known, they are confined to the crestal parts of the large anticlines of the western part of the basin. An example of one of these sandstones is shown in figure 25; this is sample S-423 of Table 4. The fossils include brachiopods, bryozoans, crinoids, probably algae, and several unrecognized forms. Locally, as on the crest of the Sholom Alechem anticline, the fossil fragments are the dominant constituents and form a small reef or biostrome. In all of the Springer sandstones that have appreciable fossil fragments the sand grain size is coarser than average, and the several coarsest Springer samples seen were all of this type.

Oolites are a minor feature but occur with some of the sandstones of the western part of the basin. They are dark colored, almost black, and are characterized by the relatively large quartz grains they have as nuclei (figure 26).

Authigenic calcite and siderite occur in a few Springer sandstones as a chemical cement. Of the two, siderite is the more abundant; but it is found mostly in the shale and shaly sands. It occurs both as scattered rhombs, so-called granular siderite (figure 27), and as large masses interbedded with the sandstones.

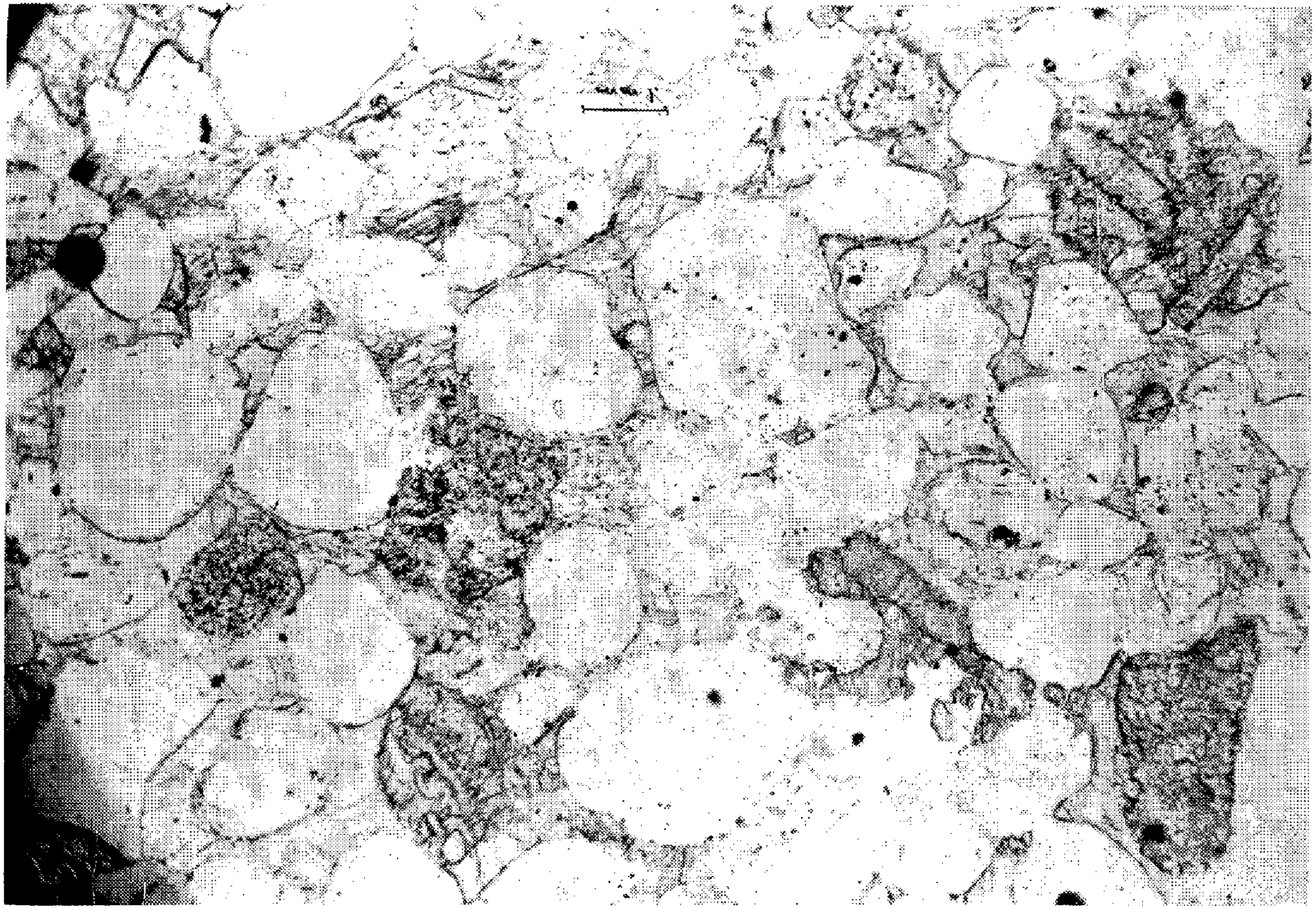


FIGURE 22. Thin-section of Hoxbar sandstone (H-Z-90). Note detrital carbonate grains. Plane polarized light.

Of the minor constituents, chert and feldspar are present in nearly every sample but usually in amounts of less than one per cent. The feldspar includes microcline, sodic plagioclase, and orthoclase. The orthoclase is badly weathered, but the microcline and plagioclase are fresh though partial calcite replacement of microcline is common.

Glauconite is prominent in some of the Springer quartzites, especially the coarser ones. In these there may be up to five per cent glauconite. None was seen in any of the other formations.

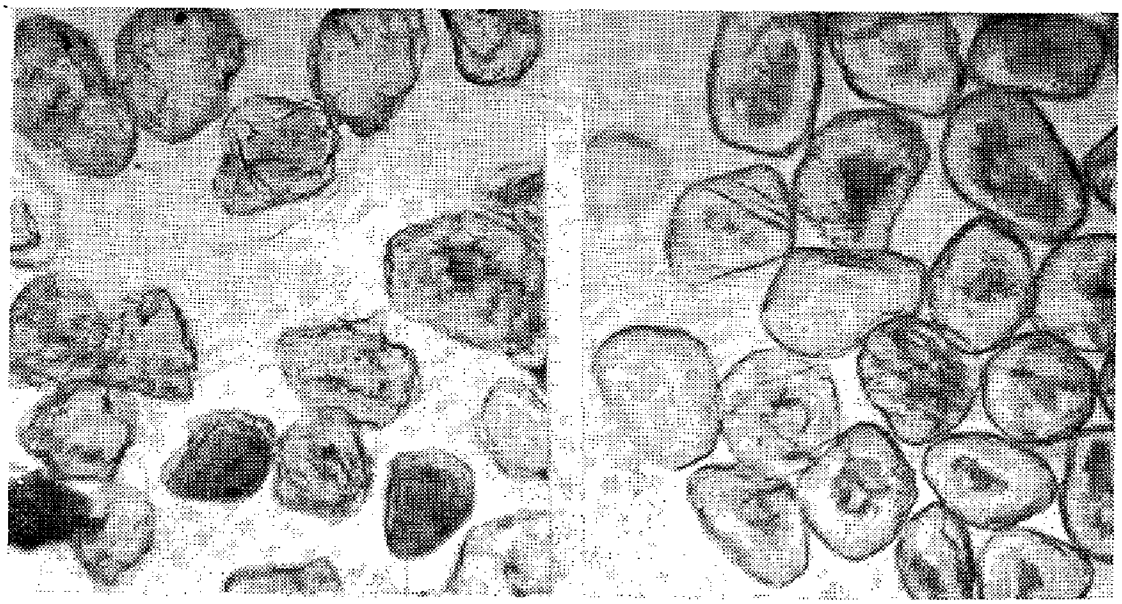


FIGURE 23. Quartz grains from sample D-70 (plane polarized light).
FIGURE 24. Quartz grains from Hoxbar sandstone (H-Z-90).

Quartzose graywackes (*rock fragment rich*).—The bulk of the Dornick Hills and Deese sandstones, the subsurface Hoxbar sandstones, and a few beds in the surface Hoxbar are rocks intermediate in composition between orthoquartzites and graywackes. This general rock type makes up close to two-thirds of all sandstones in the basin.

The median quartz content of these rocks is eighty per cent, and the range is from about 55 per cent (D-234) to 90 per cent (D-176). The balance of the sandstone is mostly various kinds of rock fragments, but also consistently present are minor amounts of feldspar, chert, and clay minerals.

The quartz in these sandstones is identical to that of the orthoquartzites in the Springer, Dornick Hills, and Deese groups

It is angular and pock-marked, and an appreciable proportion (5 to 20 per cent) is vein quartz as in the quartzites.

Rock fragments make up 2 to 32 per cent of these sandstones, and are the distinctive feature of them. The rock fragments have within any given sample a slightly coarser size distribution than the quartz grains, but in general the proportion of rock fragments is lower in the coarser-grained sandstones.

About two-thirds of the rock fragments are fine-grained clay rocks, mostly shale but possibly including some slate. Two general types can be recognized. One is gray to greenish-gray and is

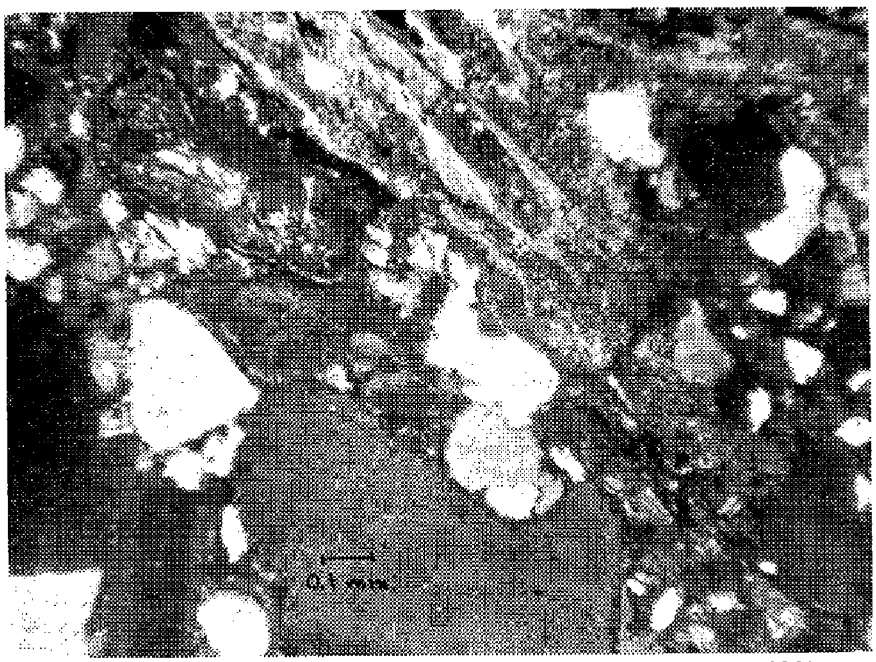


FIGURE 25. Fossil fragments in a Springer sandstone (S-S-423).

extremely fine-grained, in which no individual constituents can be determined, and there is no evidence of bedding. The second type is gray-brown, and the illite-sericite fraction is coarse enough to give a characteristic yellow to red birefringence; orientation of the fine micas is variable, but in some grains is good enough to suggest the rock may be slate rather than shale. The shale fragments are nearly all well-rounded, but an appreciable proportion, particularly of the second type, are strongly elongate. There has not been strong deformation of these soft rock fragments around the quartz grains though isolated grains warped in this fashion can be found in most samples.

Siltstone or silty shale fragments are found in nearly all samples, but they are never abundant. They consist of small quartz grains, 20 to 50 microns in diameter, distributed through a clayey matrix. The proportion of grains to matrix is somewhat variable, but the grains usually dominate.

Several varieties of metamorphic rock are represented in the rock fragments. These include phyllite, marked by its wavy bedding and relatively coarse mica, and quartz-mica schist. Probably several types of schist are present. The most common is a relatively coarse-grained (about 0.1-0.2 mm.) quartz-sericite rock in which there is a low degree of orientation. The second type has fine-grained quartz and well-developed muscovite flakes which are perfectly oriented.

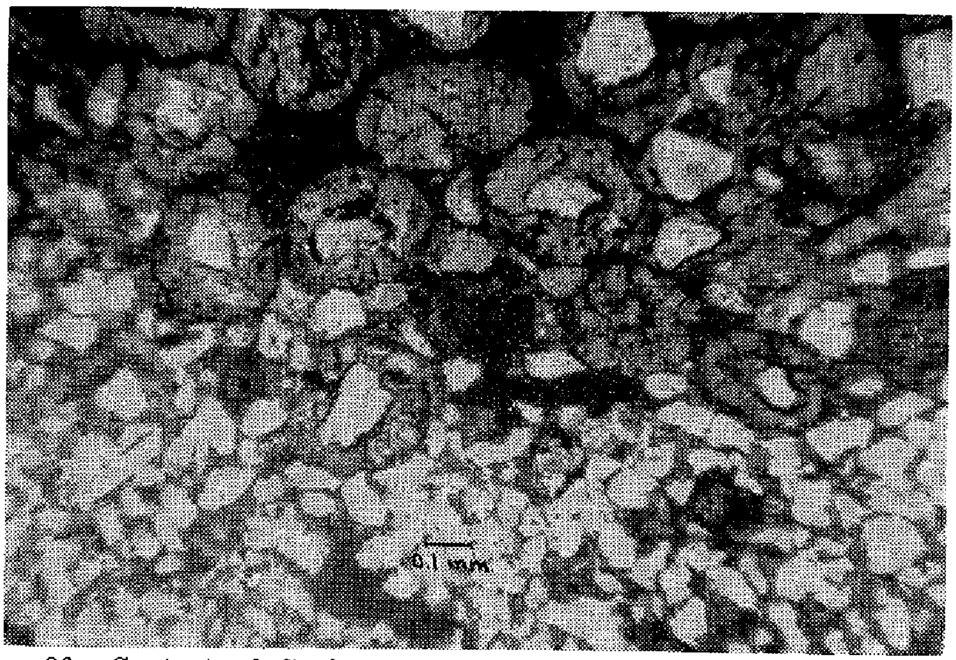


FIGURE 26. Contact of Springer sandstone and oolitic limestone (S-444). Plane polarized light.

Limestone fragments occur in most of the core samples from the western part of the basin, and constitute as much as eight per cent of some samples, but none was seen in any of the surface samples. It is possible that recent leaching at the outcrop is responsible for the absence of the limestone fragments.

Single detrital grains of gypsum were found in three samples.

Feldspar is a minor constituent of the sandstones although there is slightly more feldspar in the Dornick Hills and Deese groups than in the Springer or Hoxbar. However, the maximum amount found was only 4.4 per cent. Lucas (1934) has reported eleven per cent of feldspar from one Hoxbar sandstone, but this must be a very local feature. The feldspar is dominantly plagioclase, but includes microcline and orthoclase. The plagioclase covers some range in composition but as far as could be determined is mostly oligoclase or andesine. The degree of weathering is extremely variable; a goodly proportion, including most of the microcline and some of the plagioclase, is essentially fresh while the orthoclase and part of the plagioclase is completely clouded by kaolin or sericite.

Chert grains ordinarily do not make up more than one per cent of the sandstones. In the few samples in which chert is more abundant, the chert is present as pebbles.

A few grains of biotite were found in several of the Deese sandstones of the western part of the basin. The biotite is relatively fresh, brown, and distinctly pleochroic.

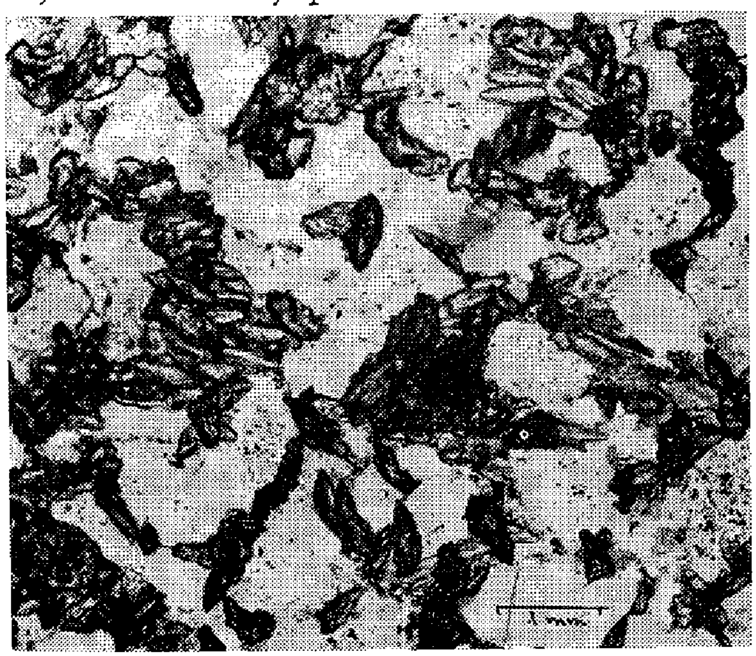


FIGURE 27. Siderite crystals in a Springer sandstone (S-H-401). Plane polarized light.

A clayey matrix is present in all samples seen, and the amount ranges from about one per cent to eight per cent. The clayey material seems to be identical to that in the shale fragments, and it may be that the matrix is merely broken-up rock fragments. The composition of the matrix could not be determined separately from the rock fragments as the fragments broke during disaggregation. However, X-rays of the fine-grained material showed illite-sericite to be the dominant constituent, and kaolinite to be present in amounts up to an estimated 25 or 30 per cent.

Clayey sandstones.—Most of the Springer sandstones are mixtures of quartz and clay. They are thus accumulations of stable minerals, but in most occurrences the clay is so abundant the rock cannot be classified as a quartzite.

The true Springer quartzites are merely the end member development of a range in mixtures of quartz and clay. Gradation from the polar quartzose sandstones is first marked by the presence of thin shale laminae or partings less than a millimeter thick. These may be separated by as much as several feet of massive

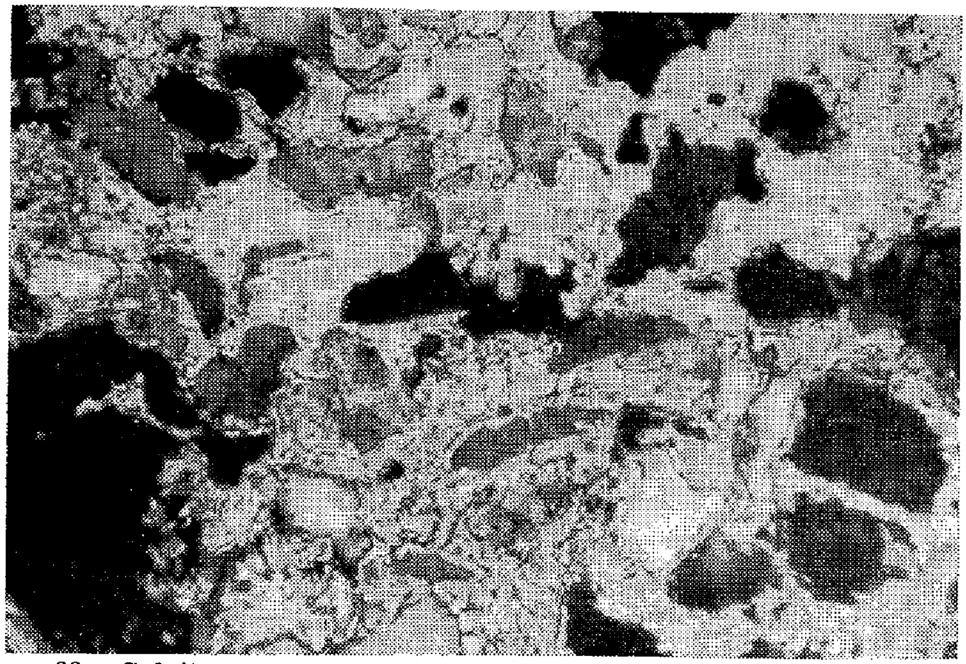


FIGURE 28. Calcite cement replacing quartz grains (S-48). Cross nicols.

quartzose sandstones, but the gradation to more clayey sands can be followed in samples which show an increased frequency and thickness of such partings. Figure 30 shows an example of one of these intermediate sandstones.

At the other end of the series there are some Springer sandstones which show no separation of clay and quartz into well-defined beds, but rather the clay occurs as a matrix between the quartz grains. The best examples of this type are found in the outcropping Rod Club member of the Springer. The clay and quartz are intimately intergrown, giving the rocks a very high degree of cohesiveness when they are unweathered (figure 31).

The quartz which occurs in these mixed rocks is no different from that in the orthoquartzites except that the average grain size is less.

The "clay" is a complex mixture that includes illite, kaolinite, montmorillonite, and chlorite. Illite is the dominant clay mineral, and it and kaolinite are present in all samples. Montmorillonite is common though never abundant, but its presence is important in the effect its swelling properties may have on oil production and oil well drilling. Chlorite was identified postively in only one sample; but as it is difficult to distinguish small amounts of chlorite when it occurs with montmorillonite, chlorite may be fairly widespread.

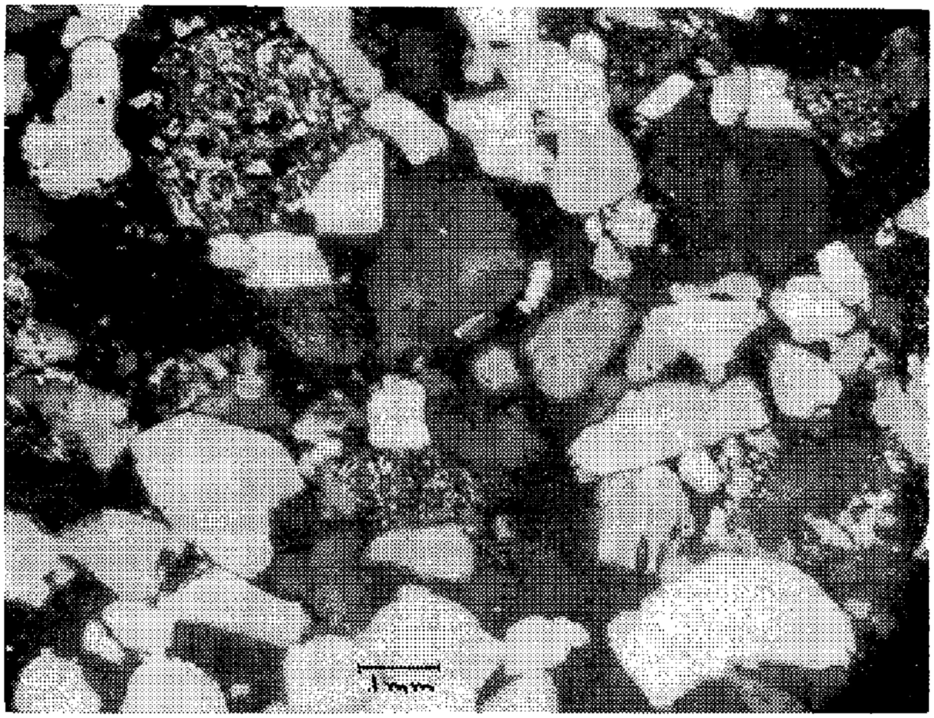


FIGURE 29. Deese sandstone containing abundant rock fragments (D-384). Crossed nicols.

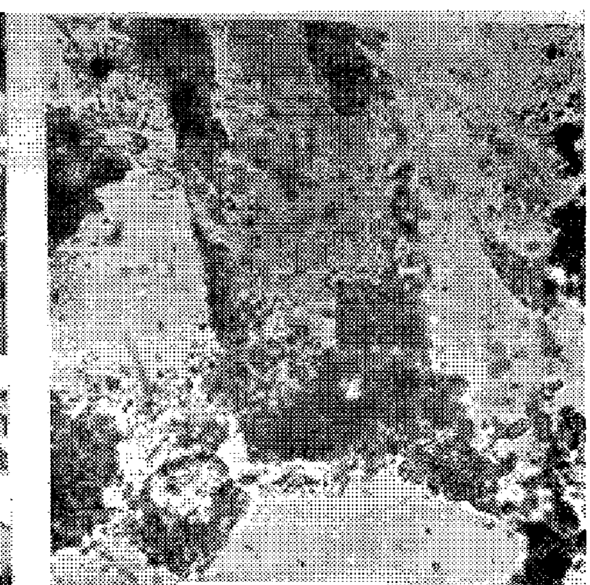


FIGURE 30. Clayey Springer sandstone with shale laminae (S-464). Crossed nicols.

FIGURE 31. Clay possibly developed authigenetically at the expense of quartz. Plane polarized light.

MINERAL COMPOSITION

(accessory minerals)

Technique.—Heavy minerals were separated by centrifuging in bromoform (S. G. 2.85) for fifteen minutes at 1,000 RPM. A sample of about two grams in approximately twenty-five cubic centimeters of bromoform was used. Separation was completed by freezing the bromoform in the lower part of the centrifuge tube and then decanting the unfrozen bromoform which contained the light fraction. All samples and separates were weighted to the nearest 0.2 milligram.

This can apparently be a very precise technique. To get a rough check on the degree of precision, replicate separations were made on five pairs of subsamples; and the mean difference in the percentage of heavy minerals was only 0.06. This variation is considerably less than the mineral counting error (Krumbein and Rasmussen, 1940); and on the basis of this limited experiment the larger variations in percentages of heavy minerals shown in Table 5 may be considered statistically significant.

The samples were split with a microsplitter from sieved size fractions. As it is impossible to escape variations in heavy mineral frequency with grain size (that is, the lighter heavy minerals will be preferentially concentrated in the finer-grained sandstones), there is no advantage in making the separate from the total sample; and it is much more convenient to work with a limited size range. Preliminary separations were made of all half-phi size fractions from a few selected samples, and from these it was found that the bulk of the heavy minerals were in approximate hydraulic equivalence with the light fraction. As nearly all of the non-opaque heavy minerals should have a size distribution one-quarter phi to one phi finer than quartz, separations of this size range should catch the minerals of greatest interest. Since most of the Ardmore basin sandstones have mean diameters between 2.75 and 3.25 phi, separations for most samples were made from the 3.0 to 3.5 phi and 3.5 to 4.0 phi size fractions. For a few of the coarser sandstones separations were also made of 2.5 to 3.0 phi fraction.

For determination and frequency counts of the heavy minerals the heavy separates were mounted in balsam or Technicon mounting medium (N=1.65). Frequency determinations were made by identifying and counting all grains passing the cross-hair intersection on equally spaced mechanical stage traverses. All minerals were counted until a total of a hundred had been reached; then the opaque and the micaceous minerals (including chlorite) were omitted. The remaining minerals were then counted until a total of 100 of these were determined. In Table 5 both the opaque minerals and the micas have been grouped, and the frequencies of these are expressed as percentages of the whole; but the remaining minerals are given as percentages exclusive of the opaque minerals and micas. In such a count and mode of expression the precision of each estimation is that of any percentage determination of a large population based on a count of 100 individuals. Figure 21 is a graph of the 95 per cent confidence interval for such an estimation; this graph is applicable to the mineral percentages of Table 5.

TABLE 5

HEAVY MINERALS OF THE ARDMORE BASIN SANDSTONES

Sample Number	Size Fraction (Phi)	Percentage of Heavy Minerals	Opaque Minerals	Mica	Tourmaline	Zircon	Rutile	Garnet	Staurolite	Apatite	Chloritoid	Clinozoisite
					Springer Outcrop							
S-RC-45	3.5-4	0.61	66		47	48	4	1				
G-165	3-3.5	0.11	64	2	83	13	1		3			
G-165	3.5-4	0.84	68	2	55	37	6		2			
G-198	3.5-4	0.58	73	1	71	20	8	2				
S-LA-174	2.5-3	0.17	86	4								
S-LA-174	3-3.5	0.09	70	4	80	10		8	2			
S-LA-174	4-4.5	2.23	74		36	55		1	2			
S-O-178	3.5-4	0.92	74		57	34	4	4		1		
S-O-242	2.5-3	0.15	65	1								
S-O-242	3-3.5	0.20	65	2	62	33	1	2	2			
S-O-242	3.5-4	0.69	60		23	64	6		6	1		
S-O-242	4-4.5	4.18	55		8	80	9		2			
					Springer Cores							
S-104	3-3.5	0.29	58		74	19	1	2		4		
S-104	3.5-4	0.83	49		45	50	5					
S-402	3.5-4	0.29	67	1	59	28	7	1		5		
S-429	3.5-4	1.42	57		36	52	6	6				
S-437	3-3.5	1.41	54		23	76	1					
S-437	3.5-4	5.09	53		16	83	1					
S-444	3.5-4	1.10	53	1	37	58	2	4				
S-457	3-3.5	0.47	61	5	84	16						
S-457	3.5 4	0.94	63	1	33	60	5	2				
					Dornick Hills Outcrop							
DH-36	3.5-4	1.98	52		18	41	11	26	1	3		
DH-B-43	3-3.5	0.24	44	4	73	13	2		11	1		
DH-B-43	3.5-4	4.41	53		29	60	6		4	1		
DH-140	3.5-4	2.05	35	2	16	17	5	56		6		
DH-B-146	3-3.5	1.50	31		10	77			13			
DH-186	3.5-4	0.57	61	9	31	44	15	6	5			
DH-220	3-3.5	0.23	73	2	72	17	9		2			
DH-240	2.5-3	0.30	84		30	25			45			
					Deese Outcrop							
D-DK-29	3-3.5	0.65	66	2	37	55	6		2			
D-32	3-3.5	0.19	52		40	12	4	39				
D-32	3.5-4	0.98	54		13	46	4	35	2			
D-32	4-4.5	0.97	40		4	54	8	32	2			
D-N-98	3-3.5	0.17	64	2	74	8	1	8	4			
D-N-98	3.5-4	0.28	74	8	43	30	5	18	4			
D-N-98	4-4.5	0.56	59	2	16	28	10	41	1	2		

TABLE 5 (Continued)

HEAVY MINERALS OF THE ARDMORE BASIN SANDSTONES

Sample Number	Size Fraction (Phi)	Percentage of Heavy Minerals	Opaque Minerals	Mica	Tourmaline	Zircon	Rutile	Garnet	Staurolite	Apatite	Chloritoid	Clinozoisite
D-130	3-3.5	1.28	76	1	52	31	9	4	3	1		
D-151	3-3.5	0.27	62	2	51	25	2		21			
D-151	3.5-4	0.52	57	2	34	50	7		8			
D-176	2.5-3	0.59	74	1	78	4	2		6	10		
D-176	3-3.5	3.72	56	1	11	73	8	3	4	1		
D-176	3.5-4	0.52	52		15	76	4	2	3	1		
D-181	3-3.5	0.55	66	3	54	20	5	14	7			
D-DK-214	3-3.5	0.50	61		58	33	5	2	2			
D-DK-214	3.5-4	0.54	45	2	38	48	7		7			
D-DK-214	4-4.5	0.50	50		10	73	8	1	8			
D-218	3-3.5	0.18	67	4	56	14	4	18	9			
					Deese Cores							
D-70	3-3.5	0.63	40	3	32	4	2	59		2	1	
D-278	3-3.5	0.95	48	4	14	8	6	67		1	4	
D-278	3.5-4	3.22	40	2	9	36	2	53				
D-284	2.5-3	0.19	64	9	33	34	2	26	2	2	1	
D-284	3-3.5	0.20	59	10	38	10	3	42	3	3	1	
D-284	3.5-4	0.55	51	3	25	24	2	40	1	4	1	3
D-284	4-4.5	1.72	43	1	14	26	7	45	2	2	2	2
D-286	3.5-4	0.27	80	7	51	36	4	3	2		4	1
D-305	3.5-4	0.42	65	1	18	68	11	1		1	1	
					Hoxbar Outcrop							
H-34	3-3.5	0.32	72	2	57	10	4	23	5			
H-84	3-3.5	0.24	56	3	27	35	9	24	5			
H-84	3.5-4	0.38	52	2	14	63	10	10	3			
H-Z-90	3-3.5	0.17	17		30	59	2	9				
H-Z-90	3.5-4	0.31	28		26	58	3	8	5			
H-Z-90	4-4.5	0.35	34		21	59	6	12	2			
H-236	3-3.5	0.14	46	4	57	4	5	18	15			1
H-236	3.5-4	0.31			39	31	7	16	4			3
H-239	3-3.5	0.10	68	4	54	34	1	5	6			
H-239	3.5-4	0.34	60	2	16	70	2	5	5	1	1	
					Hoxbar Core							
H-290	3.5-4	0.76	53	7	26	14	3	46	2	4	3	2

Description of minerals.

Anatase. Several grains of a dark steely blue corroded and worn highly birefringent mineral were tentatively identified as anatase. These grains are detrital.

Apatite occurs sporadically throughout the samples but nowhere comprises more than a few per cent of the non-opaque minerals. It is perhaps the most erratic in its occurrence of all the heavy minerals. It occurs as both well-worn rounded grains and as irregularly broken grains. The grains are colorless and contain few inclusions. A few grains had stringers of small cavities oriented parallel to the "c" crystallographic axis.

Barite is present in nearly all samples which contain carbonate and in several is present in flood amounts. The barite grains are extremely angular and irregular in shape, probably reflecting the occurrence of the mineral as a cement. Inclusions are abundant and are mostly irregular vacuoles. Undoubtedly the barite is authigenic.

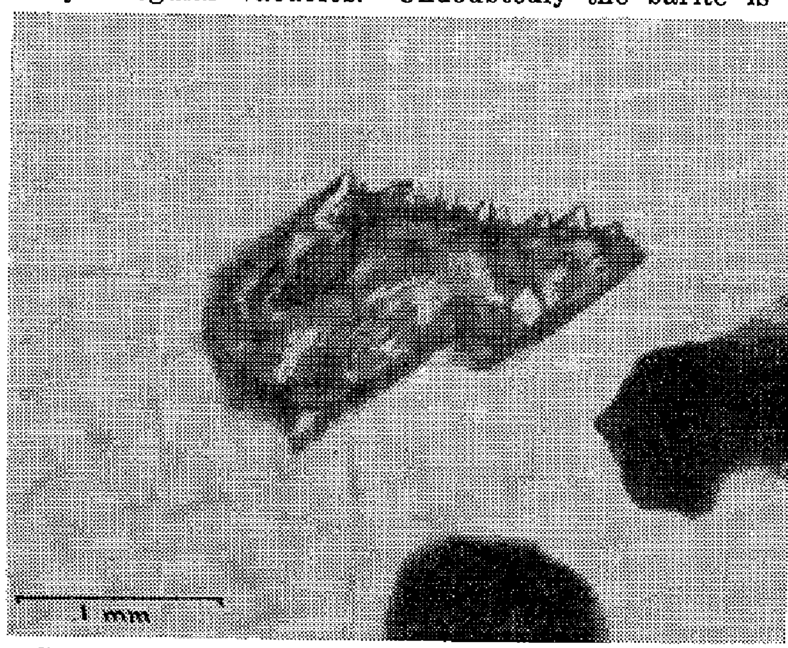


FIGURE 32. Serrate staurolite grain from Dornick Hills sandstone (DH-B-43). Plane polarized light.

Chlorite. Chlorite group minerals occur in two distinctive forms. One is dull gray-green with no recognizable pleochroism and shows splotchy uneven birefringence; it occurs in very irregular flakes and is found in small amounts in nearly all samples but is most abundant in the finer-grained sandstones. The second type of chlorite is a clear moderate green and has a distinct pleochroism. It occurs in well-defined flakes, some of which have abundant rutile inclusions. This chlorite is similar to that characteristically found in metamorphic rocks of moderate rank. In the Ardmore sediments it is a minor constituent; it was found in the Deese and Hoxbar sandstones from the western part of the basin, but only in the Hoxbar sandstones from the eastern part.

Chloritoid. The distribution of chloritoid is the same as that of the second type of chlorite described above; it is never abundant, and where present makes up from 1 to 5 per cent of the non-opaque minerals. It occurs as platy grains, has the characteristic green-blue pleochroism, and extremely low birefringence. A few grains had an imperfect cleavage at an angle of 70 degrees with the more conspicuous prismatic cleavage.

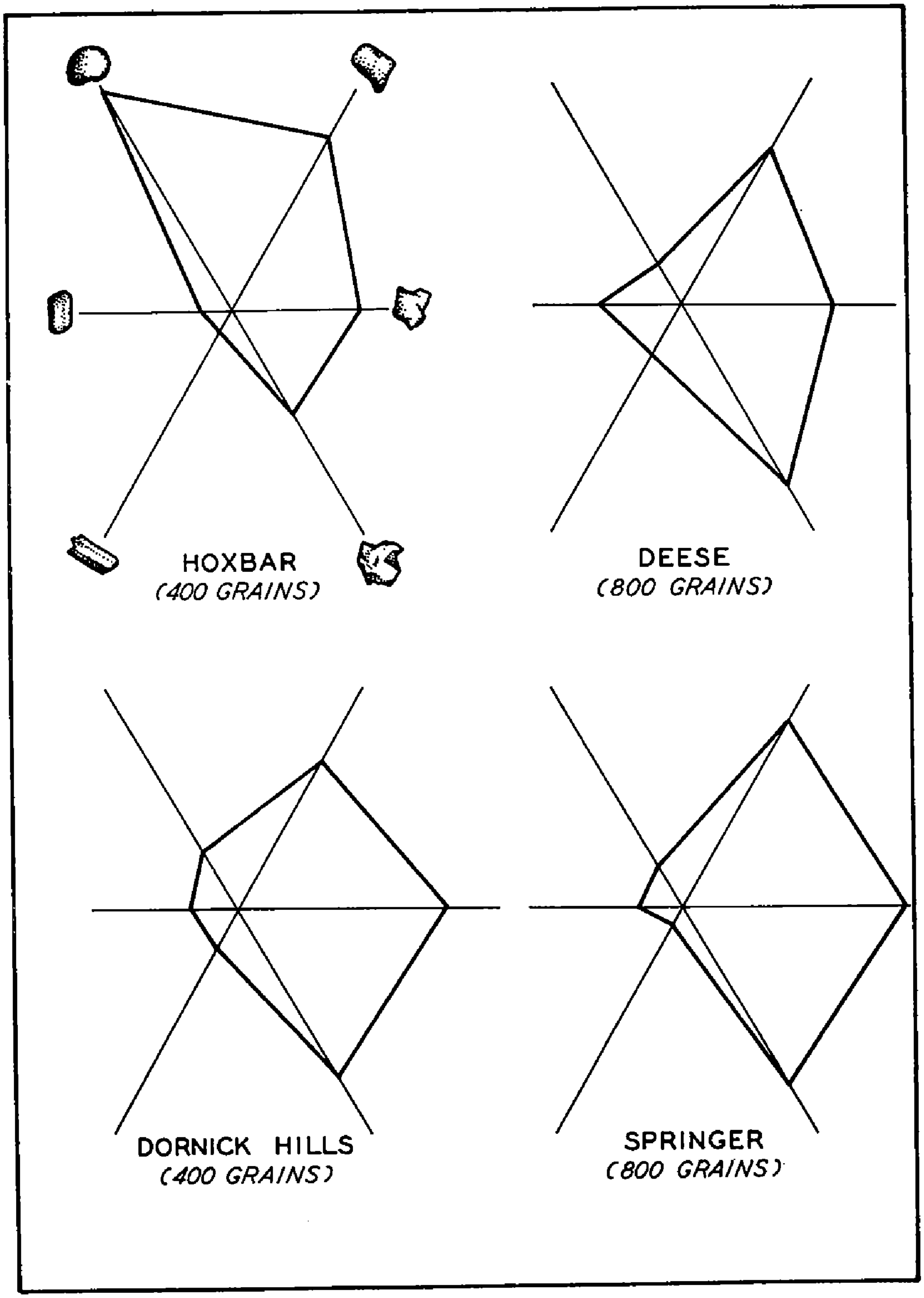


FIGURE 33. Shape of tourmaline grains.

Clinozoisite. A few grains of clinozoisite are present in most of the core samples from the Deese and Hoxbar formations. It is present as subangular nearly equant grains which are identified by low birefringence and anomalous uneven yellow-gray interference colors. Apparently some range in composition is present as about half the grains are colored a pale greenish-yellow. These colored grains are distinctly pleochroic. Several of the grains had a pitted surface similar to that which is commonly found on garnet.

Collophane. Phosphate minerals are present in small amounts in nearly all samples. Many grains have some indication of an internal structure which is probably organic in origin; these grains typically are not entirely isotropic but show irregular low birefringence.

Epidote. A few greenish-yellow pleochroic highly birefringent grains occur with the clinozoisite.

Garnet. Although very erratic in its occurrence garnet is one of the more abundant heavy minerals, and is the dominant mineral in many separates. Garnet is a very minor constituent of the Springer sandstones, but is abundant in most of the others (Table 5). It is present entirely as angular to subangular grains, many with marked conchoidal fracture. A small proportion of the grains (less than 5 per cent) have part of their surface irregularly pitted, but there are no skeletal forms which might suggest intrastratal solution, nor is there any evidence of alteration of the garnet.

The range in color is from colorless to pink, and there appears to be a complete gradation. Most grains are colorless, and distinctly colored grains are rare.

The garnet does not seem to be in hydraulic equivalence with the light minerals. Theoretically, if availability is not a factor, the frequency of a mineral in any size fraction is dependent upon the mean size and degree of sorting of the total sample (Rittenhouse, 1944); and hence, there should be a high degree of correlation of mineral frequency with size and sorting. Table 6 lists the garnet frequency in the 230 mesh size fraction from nine samples, and the mean size and standard deviation of the samples. The correlation has been tested by the method of multiple correlation (Ezekiel, 1941). Also given is the equation of best fit. The correlation coefficient is very low (0.43) and is not statistically significant. When corrected for the number of samples, it actually becomes 0.0. This lack of correlation is considered strong evidence that the distribution of the garnet has been controlled by the source area and, further, that this is the first cycle of sedimentation for the garnet.

Ilmenite. This is the most abundant of the opaque minerals, and is probably the most abundant of all heavy minerals. It occurs as subrounded equant grains. It is a major constituent of all samples, but is slightly less abundant in the Hoxbar samples of the outcrop area.

Leucoxene. Leucoxene is abundant in all samples. Many grains are translucent along thin edges and show high birefringence; and the characteristic dull white color in reflected light is frequently stained yellow, presumably by limonite.

Magnetite. Magnetite is much less common than ilmenite, but its presence in small amounts in most samples can be demonstrated with a magnet.

Pyrite. Pyrite is a common accessory mineral in the subsurface samples; but is absent in all outcrop samples, presumably having been destroyed by weathering. The pyrite occurs as grain size nodules made up of a large number of intergrown crystals. These might be termed "micro-concretions", and are clearly authigenic.

Rutile. Rutile is present in all samples; and, as expected from its density, is most abundant in the finer grain sizes. Somewhat abraded prismatic grains are most common, but well-rounded grains are found in samples with well-rounded tourmaline and zircon. In

TABLE 6
RELATION OF GARNET AND TOURMALINE FREQUENCIES (230 MESH) TO GRAIN SIZE FREQUENCY DISTRIBUTION

Sample No.	Mean grain size	Standard deviation	Per cent garnet	Per cent tourmaline
D-176	2.54	0.19	2	15
D-278	2.74	0.32	53	9
DH-36	2.93	0.25	26	18
D-32	3.00	0.54	35	13
DH-140	3.04	0.26	56	16
D-284	3.12	0.74	40	25
D-N-98	3.15	0.43	18	43
DH-186	3.19	0.33	6	31
D-286	3.58	0.32	3	51

Multiple correlation of garnet (X_1), mean grain size (X_2), and grain size standard deviation (X_3)

$$X_1 = 83.44 - 24.36\ X_2 + 45.20\ X_3$$

$$R = 0.43 \qquad \bar{R} = 0.0$$

Multiple correlation of tourmaline (X_1), mean grain size (X_2), and grain size standard deviation (X_3)

$$X_1 = -99.75 + 42.53\ X_2 - 12.38\ X_3$$

$$R = 0.84 \qquad \bar{R} = 0.77$$

a few of the less abraded grains longitudinal striations were seen.

Color is either brownish-yellow or deep brownish-red, and the two are about equally common. A slight but distinct pleochroism is present.

Spinel. Two grains of spinel were seen, one in a Springer sandstone sample and the other in a Deese sample. Both were of the light green variety commonly known as ceylonite.

Staurolite. Staurolite is erratic in its occurrence, but is abundant in some of the Dornick Hills, Deese, and Hoxbar sandstones. It occurs as irregular fragments, many showing a conchoidal fracture. Occasional grains show the effect of solution along cleavage traces by having markedly serrate boundaries (figure 32).

The color ranges from pale yellow to yellow-brown, with moderate pleochroism. Small dust-like opaque inclusions are abundant in many grains, and large vacuoles are seen occasionally.

Tourmaline is an abundant constituent of all the heavy mineral separates, and is found with a wide range of color and shape.

Two very distinctive shape assemblages are present, a well-rounded one that was found only in the outcropping Hoxbar sandstones and an angular one which occurs in all of the other formations.

The well-rounded assemblage shows extreme modification, presumably by abrasion; and almost no trace of the original shape of the grains remains. The mean roundness of 400 grains (determined by visual comparison with standard images) was 0.75; however, the shape distribution is strongly skewed, and in individual samples one-half to three-quarters of the grains have a roundness value of 0.8 or higher. In most samples with the well-rounded grains there is also an admixture of the angular assemblage, giving a bimodal shape frequency distribution.

Most of the extremely well-rounded grains are nearly spherical, but fifteen to twenty per cent are somewhat flattened and approximate oblate spheroids. This oblate shape apparently reflects the basal parting of tourmaline; and as the grains tend to come to rest on a slide with their short dimension perpendicular to the surface of the slide, these oblate grains present a basal section and are black and non-pleochroic when seen under the microscope.

Green is the most common color in the well-rounded tourmaline, and approximately 55 per cent of the grains are pleochroic in shades of light and deep green. About fifteen per cent have a pleochroic formula of yellow-brown to deep brown; seven per cent are blue with only slight pleochroism; five per cent have a dichroic scheme of pale pink and deep green; and the remainder are black and non-pleochroic.

Inclusions are not common, but conspicuous large vacuoles are present in a small proportion of the grains.

This well-rounded assemblage is like the tourmaline from the Ordovician sandstones of the Arbuckle Mountains (figure 10), and is believed to have been derived from these sandstones.

The angular assemblage of tourmaline displays extreme variation in both color and shape. The following color varieties can be distinguished:

1. Brown tourmaline

   a. Pleochroic in various shades of faint yellow to deep brown. This color occurs mostly in small stubby prisms, and as a result is most common in the finer size fractions. In any given size fraction its frequency varies with the mean size of the sample, being invariably more abundant in those with the coarser mean. In the 170 mesh fraction from ten samples the frequency of this color variety ranged from 15 to 40 per cent.

   b. Pleochroic in shades of olive-brown, pleochroism weak. Occurs mostly as relatively large angular anhedral grains. Many grains are elongated due to fracturing along basal parting planes and, hence, show an apparent reversal of the normal tourmaline pleochroism with maximum absorption parallel to the prismatic elongation. This color variety is characteristically free of inclusions. It is about one-half to two-thirds as abundant as the yellow-brown type.
2. Green tourmaline with pleochroism in shades of light green to very dark green makes up about 35 per cent of the angular tourmaline assemblage. It occurs both as small prisms and as irregular fragments.
3. Blue tourmaline

   a. Pleochroic from nearly colorless to gray-blue. This is a relatively rare type and occurs only as small prisms.

   b. Moderate blue, virtually non-pleochroic. In contrast to the gray-blue variety this never shows idiomorphism. It is also uncommon.

   c. A few grains of blue tourmaline with a closely spaced network of rectangular fractures parallel and perpendicular to the prism were seen. The origin of the fractures is not known, but the grains are distinctive.
4. Pink tourmaline, pleochroic from various shades of pink to dark green, is a common variety. It occurs mostly as slightly modified short prisms.
5. Varicolored tourmaline, usually a combination of olive-brown and blue, is found in an occasional large fragmental grain.
6. Authigenic tourmaline overgrowths are found on two to three per cent of the grains. This tourmaline is colorless or very light blue.

The tourmaline shows a high degree of shape sorting; and, hence, there is considerable variation of average shape with the mean grain size of the sample from which the separation was made and with the size fraction. However, the mean roundness of 500 grains from ten samples was 0.45. Figure 33 illustrates the variation in shape in another way. The grains were divided into size shape classes, rounded, subrounded, subangular, angular, rounded prisms, and angular prisms. The proportions of grains in the several classes were then plotted on the sailboat diagrams of figure 33. It can be seen that the variation between formations is slight except for the Hoxbar formation. This reflects the presence of the rounded assemblage in the Hoxbar samples.

Inclusions in the tourmaline are mostly vacuoles, some of which are oriented parallel to the prism; but also found are zircon, rutile, and fine opaque dust which could be carbonaceous matter or iron oxides. The inclusions are most common in the prismatic grains, and the large fragmental grains are characteristically clear.

Both types of tourmaline occur in approximate hydraulic equivalence to quartz and, hence, are most common in the half-phi interval just finer than the sample mean. The frequency of tourmaline in a given size fraction shows a high correlation with mean sample size and sorting. Table 6 shows the tourmaline frequency in the 230 mesh fraction, the mean size, and the standard deviation for the same samples in which the garnet correlation described above was attempted. In contrast with garnet the multiple correlation of tourmaline frequency with size and sorting is high, with a multiple correlation coefficient of 0.78 (reduced for number of samples). This is significant at a 0.01 level, and indicates that the frequency of tourmaline is more a function of the sample size characteristics than of availability.

TABLE 7
SHAPE OF ZIRCON GRAINS
(230 MESH, n = 25)

	axial ratio		roundness	
Sample	X	s	X	s
S-RC-45	0.699	0.126	0.55	0.10
S-O-242	0.658	0.144	0.58	0.088
S-444	0.665	0.143	0.53	0.11
S-457	0.672	0.120	0.60	0.091
DH-36	0.676	0.118	0.50	0.094
DH-186	0.658	0.125	0.52	0.099
D-151	0.642	0.119	0.52	0.109
D-DK-214	0.705	0.139	0.54	0.12
D-278	0.634	0.166	0.52	0.10
D-305	0.698	0.143	0.47	0.10
H-Z-90	0.670	0.150	0.68	0.15
H-290	0.647	0.155	0.52	0.090

Zircon. The mineral zircon is present in relatively large amounts in all samples, and is probably the most abundant of the non-opaque minerals. In all samples it appears to be in hydraulic equivalence to quartz, and its greatest frequency is in the size-fraction approximately one phi unit finer than the sample mean. It is in nearly all samples the dominant mineral in the finer-grained separations (Table 5).

Like tourmaline, zircon occurs in both a well-rounded and angular shape assemblage. The well-rounded zircon occurs with rounded tourmaline in the Hoxbar sandstones; and though not as well rounded as the tourmaline, it is distinctly better-rounded than the zircon within the older formations.

The more angular zircon of the Springer, Dornick Hills, and Deese formations occurs mostly as anhedral fragments; but in each sample, particularly in the finer sizes, a few euhedral or subhedral grains are present. The crystal habit is the normal one of prism and pyramids.

Table 7 lists the axial ratio (b/a) and estimated roundness of 12 samples of 25 grains each. It is perhaps surprising to find the mean axial ratio to be approximately the same as that for quartz (Table 9). The crystal habit of zircon is notably elongate, and it might have been expected that more pronounced elongation would be inherited by the detrital zircon.

Most of the zircon is colorless, but yellow to brown zircon is common, and a few grains of pink to purple varieties can be found in nearly every sample.

Inclusions occur in about one-third of the grains. Microlites of euhedral zircon and rutile are most common, but also present are opaque inclusions and vacuoles. There is no tendency for the inclusions to be oriented. A few grains showed an opaque rounded central "kernel", the character of which could not be determined.

Zoning is generally uncommon, and is mostly confined to the larger grains. Where present, the "zones" are exceedingly numerous and fine.

In a few grains anastomosing cracks, similar to those described by Hunton (1950), were seen. These grains had abnormally low birefringence.

Heavy mineral assemblages.

In a general way the heavy minerals of the Ardmore basin sandstones are relatively simple. Only relatively stable minerals are present, and the number of species is not large. However, several distinctive assemblages are present; and each of these is limited to a definite stratigraphic interval.

The Springer sandstones carry a non-opaque assemblage which is dominated by the ultra-stable minerals tourmaline, zircon, and rutile but with small amounts of garnet, apatite, and staurolite. Large amounts of ilmenite and leucoxene are also present. All of these minerals occur as markedly angular grains. This is a typical assemblage of heavy minerals from relatively quartzose sandstone. The dominance of the ultra-stable minerals reflects the long sedimentary history of the detritus; and the small amounts of garnet, etc., are perhaps the remnants of the more complex sediment from which the Springer sands were derived.

A notable feature of Springer sandstones is the high concentration of heavy minerals in a few samples (S-437, for example). These samples with the large amounts of heavy minerals are all from oil reservoirs in the western part of the basin, and are thought to be placer concentrates due to relatively strong current agitation during deposition . These samples also have smaller amounts of the less stable minerals.

The Dornick Hills and Deese sandstones carry the same minerals as the Springer, plus an assemblage of metamorphic minerals (figure 34) of which garnet and staurolite are the most abundant but with small amounts of chloritoid, metamorphic-type chlorite, and clinozoisite. An attempt was made to separate the tourmaline and zircon of the Springer sandstones from that in the Dornick

Hills-Deese sandstones on the basis of color, shape, and type of inclusions; but no statistically significant difference could be found in the study of 2,000 grains. However, such differences would have to be relatively large to be detected. Color varies with grain shape, and shape in turn varies with grain size. Hence, the frequencies in any shape and color classes which are erected are sensitive to minor changes in the grain size distribution of the total sample.

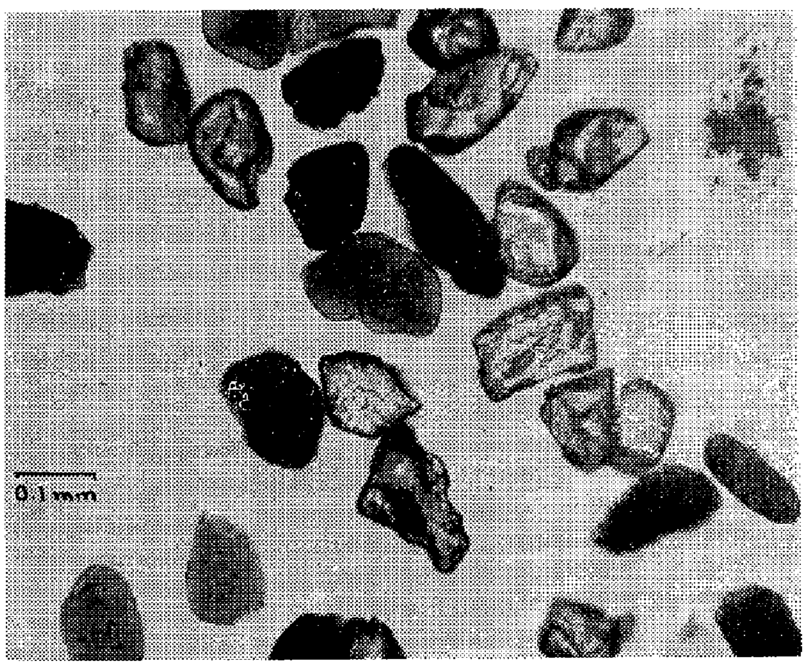


FIGURE 34. Heavy minerals from a Deese sandstone (D-284). Plane polarized light.

The metamorphic minerals are very erratic in their occurrence, and on the number of samples studied it is not possible to recognize any difference between the Dornick Hills and Deese. No chloritoid, chlorite, or clinozoisite was found in the Dornick Hills sandstones; but these minerals occur in all of the Deese samples from the western part of the basin and in the uppermost Deese of the eastern part of the basin. However, as no samples were obtained from the Dornick Hills in the west, this difference may be only the result of sampling. Sandstones which occur with the Bostwick conglomerate have an admixture of well-rounded heavy minerals similar to those in the Hoxbar sandstones.

The metamorphic heavy minerals are not in hydraulic equivalence with the light minerals though the tourmaline, zircon, and rutile apparently are. This suggests that the metamorphic minerals have had a different sedimentary history than the remainder of the rock, and are probably a contamination added to the sediments in their latest depositional cycle. The lack of hydraulic adjustment of the metamorphic minerals could conceivably be due to intrastratal solution, but other evidence (see below) suggests these minerals had a source independent of the more stable minerals.

Hoxbar samples from cores in the western part of the basin are not different from the Deese samples, but the outcropping Hoxbar sandstones have a very distinctive assemblage of well-rounded tourmaline, zircon, and rutile (figure 35). The high degree of rounding sharply differentiates this suite from the same minerals in the older formation. This assemblage is like that found in the Ordovician sandstones of the Arbuckle Mountains (figure 10) and is likely the product of the reworking of these sandstones.

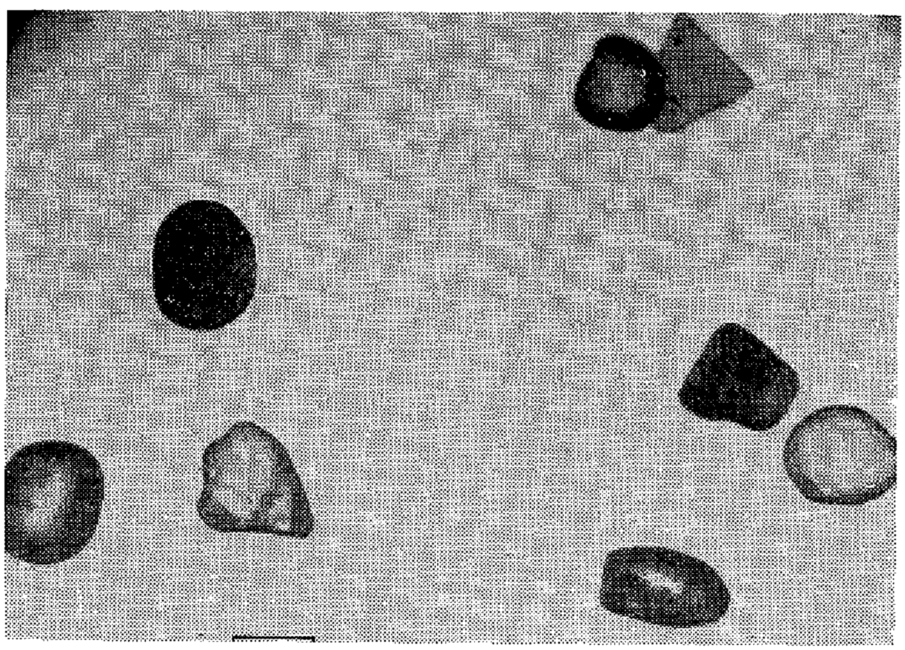


FIGURE 35. Heavy minerals from a Hoxbar sandstone (H-Z-90). Plane polarized light.

These well-rounded ultra-stable minerals occur with angular grains of the metamorphic minerals found in the Deese sandstones. This occurrence of the same suite of metamorphic minerals with two suites of ultra-stable minerals further supports the interpretation of the metamorphic minerals having a separate source.

In some of the Hoxbar samples there is also a subordinate admixture of angular tourmaline, zircon, and rutile. This fact, along with absence of the rounded assemblage in the western part of the basin, indicates the source of the rounded grains was local, which is in keeping with the Arbuckle Mountains being the uplift which was being eroded.

Grain Roundness

Technique.—For rapid work on a large number of samples the only practical technique for measuring roundness is by visual comparison of grains with standard images (Krumbein, 1941). The roundness values reported here were estimated by microscopic observation of grains from the 70 mesh sieve fraction mounted in a medium with an index of refraction of 1.65. With this mounting medium the grain boundaries show very clearly. A single sieve fraction was chosen to minimize variations dependent upon size and to allow uniform magnification to the size of standard images. The 70 mesh size was chosen because it is the coarsest fraction common to nearly all samples and because it lends itself well to magnification to the approximate size of the standard images of Krumbein's chart (25 mm).

Roundness estimations were made for 21 samples; for 13 of these roundness was determined on 50 grains; and for the other 8, on 100 grains. Grains were randomly chosen by making roundness estimations of all grains passing the cross-hair intersection during equally spaced mechanical stage traverses.

As a check on the precision of the technique, replicate analyses were made on eight samples. Roundness values of fifty grains in each of the eight samples were estimated twice with a two week interval between the determinations. The average arithmetic difference between replicates was 0.021; and the maximum, 0.040. These differences are so much less than the variation between samples that it can reasonably be concluded that differences between samples are not due to lack of precision in the technique.

Description of results.—Roundness data for the 21 samples are summarized in Table 8. The range in sample roundness is 0.326 to 0.641. Most of the samples form one population with a mean roundness of about 0.39, but three samples from the Hoxbar comprise a group by themselves with a roundness of about 0.60. The difference between the Hoxbar grains and the others is an obvious one and is illustrated by the photomicrographs of figures 23 and 24. Actually this high degree of roundness applies only to the outcropping Hoxbar sandstones; the fourth Hoxbar sample (H-68) is from the Healdton oil field and belongs with the angular group of samples.

TABLE 8
ROUNDNESS ESTIMATED BY VISUAL COMPARISON
(70 MESH SIZE)

Sample	Mean	Standard deviation	Sample	Mean	Standard deviation
	Springer			Deese	
S-78	0.326	0.0912	D-DK-29[1]	0.372	0.0869
S-165	0.370	0.0900	D-70	0.444	0.0837
S-L-190[1]	0.442	0.1116	D-N-98[1]	0.387	0.0990
S-O-243	0.402	0.0915	D-151	0.368	0.0914
S-S-437[1]	0.368	0.0926	D-218	0.370	0.0833
S-S-455	0.380	0.0833	D-272	0.366	0.0861
			D-284	0.344	0.0929
	Dornick Hills			Hoxbar	
DH-B-431	0.378	0.0886	H-68	0.440	0.0728
DH-140	0.410	0.0814	H-Z-90	0.580	0.1107
DH-B-146	0.442	0.0785	H-236[1]	0.582	0.1313
DH-B-240	0.408	0.1007	H-239[1]	0.641	0.1355

[1] 100 grains.

The mean roundness of the Springer samples is 0.381; of the Dornick Hills samples, 0.409; and of the Deese samples, 0.379. These differences are not significant as both variation between samples and variation due to experimental error is as great as the differences between formations.

Grains from sandstones other than the outcropping Hoxbar sandstones are marked not only by low degree of rounding but also by very irregular surface texture. This is illustrated in figure 23. The surface of the quartz is pitted and etched in a manner suggesting authigenic chemical attack on the quartz. As most of the samples can be seen in thin-section not to have any appreciable amount of clay matrix or chemical cement which could have been the agent of attack, it is probable the irregular surface texture has been inherited.

Conclusions.—The great difference in roundness between the Hoxbar and the other sands clearly indicates a change in the source of the sediments. The roundness of the Hoxbar sands is matched only in first or second cycle orthoquartzites, such as those in the

Simpson group of the Arbuckle mountains. The Hoxbar grains are identical to those of the Simpson, and there seems no reason to assign them to any source other than the Simpson. The fact that most of the subsurface Hoxbar does not carry the rounded grains indicates that the source was local, as it would be if the Arbuckle Mountains were the uplift supplying the rounded quartz.

Grain Shape

Technique.—As a measure of geometric form the ratio of the grain axes lends itself well to simple procedure and rigid operational definition. The ratio used herein is that between the "a" and "b" axes, in which the "a" axis is defined as the maximum grain intercept and the 'b" axis as the maximum intercept perpendicular to "a."

All measurements were made from thin sections by means of a micrometer ocular. Grains measured were selected randomly either by use of a mechanical stage equipped with a click-wheel or by measuring all grains intersected by lines ruled on the thin-section cover slip.

Description of results.—Results for the twelve measured samples are presented in Table 9 as mean axial ratio and standard deviation of the individual grain determinations. The number of grains measured is also given.

TABLE 9
AXIAL RATIOS

Sample	Mean	Standard deviation	Number
S-RC-45	0.72	0.162	25
S-402	0.66	0.160	50
S-437	0.70	0.172	25
DH-B-43	0.67	0.153	50
DH-186	0.64	0.116	25
DH-140	0.67	0.116	25
D-32	0.70	0.108	25
D-DK-40	0.72	0.149	25
D-70	0.69	0.142	25
H-84	0.74	0.111	100
H-Z-90	0.73	0.126	25
H-86	0.73	0.118	25

The range in values for mean axial ratios is only from 0.64 to 0.74. The three samples from the Hoxbar formation are slightly more equant than those from the other formations, but it is not apparent that the small differences between individual samples or between formations are significant. However, these data lend themselves to a completely randomized design for analysis of variance whereby the significance of the observed differences can be checked.

Four formations are represented, and three samples were measured in each. Twenty-five grains were measured in each sample (where more than 25 had been measured, 25 were selected at random for the analysis). Thus, there are differences between formations, differences between samples, and differences between grains. Table 10 is a summary of the analysis of variance.

TABLE 10
ANALYSIS OF VARIANCE FOR AXIAL RATIOS

Source of variance	Degree of freedom	Sum of squares	Mean square	F	F.05	F.01	P
Formations	3	0.1880	0.06267	3.05	2.62	3.82	0.01-0.05
Samples within formations	8	0.1706	0.02132	1.04	2.09		N.S.
Grains	288	5.9164	0.02054				
Pooled error	296	6.0870	0.02056				

A statistically significant difference between formations is indicated by the analysis of variance. This reflects the high degree of roundness in the Hoxbar samples.

There appears to be an inverse relationship between axial ratios and grain diameters within individual samples. Considering all grains no correlation is suggested; but in each sample, regardless of the absolute grain size, the coarser grains tend to be the most elongated. The correlation is not high and in some of the samples is not statistically significant, but it is never reversed.

The relationship cannot be attributed to the change in precision of measurement with grain size since it is independent of absolute size. That is, probabilities are that two grains of given size will have different axial ratios if one is a coarse grain in its particular sample and the other is a fine grain in its sample. The necessary conclusion is that there is an appreciable and measurable element of shape sorting within samples.

Conclusions.—Relatively little data is available with which to compare these axial ratio values. Griffiths and Rosenfeld (1951) report the axial ratios of 4,400 grains of the Bradford sand, and this is the largest reservoir of data available. They found a mean value of 0.650-0.682 and a standard deviation of 0.144-0.150, which compares with 0.698 and 0.145 for the grains measured here. Bokman (1951) found the mean axial ratio for the Stanley formation to be 0.645 and for the Jackfork formation to be 0.585, but the writer has not been able to duplicate these results although using the same technique. Available information (Curray, 1951) suggests that axial ratios of quartz grains vary over only a very small range in different rock types, and that most samples will fall between 0.65 and 0.70. Despite the small amount of available data any departure from this range may be considered unusual, and it is probably justified to conclude that the Hoxbar is an unusually equant-grained sand.

Grain Size

Of the numerous methods available for measuring grain size, sieving has been generally considered the simplest and most practical for material of sand size and, insofar as possible, has been used in this study. However, several samples for which size analyses were desired were core fragments and were too small for sieving, and a few samples were so fine-grained it was thought sieving was not practicable; and for these, size distributions were determined from thin sections.

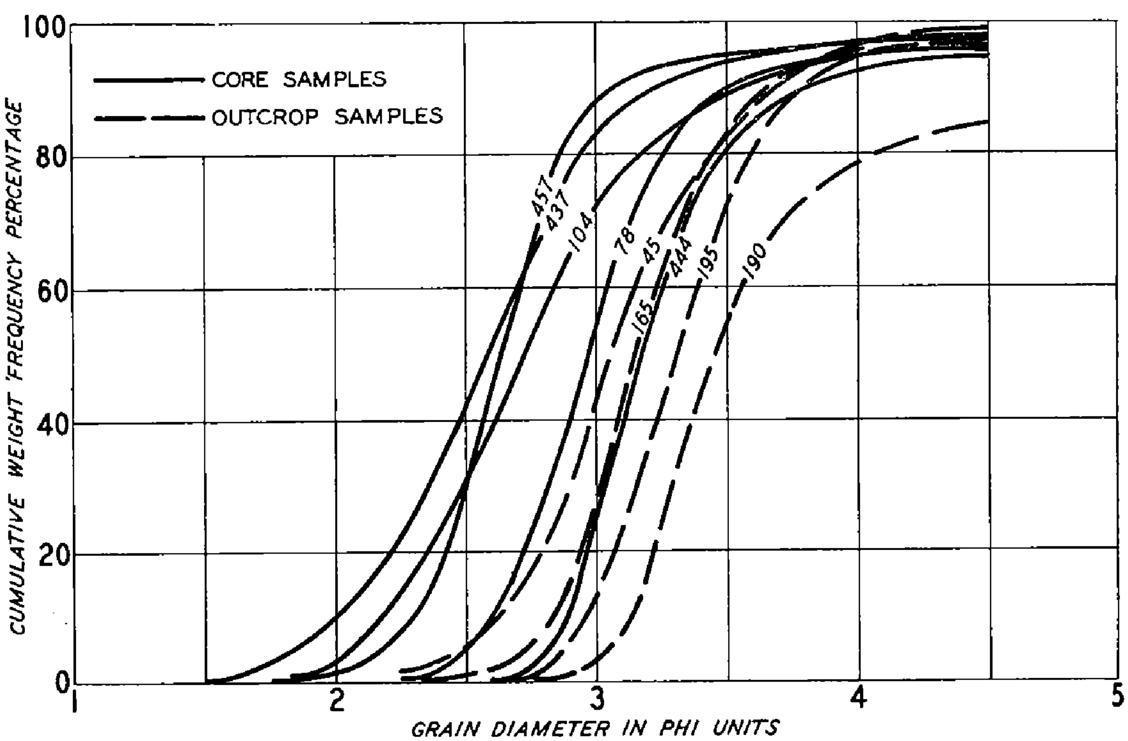


FIGURE 36. Cumulative size frequency distributions of Springer sandstone samples.

Sieving Technique. For the most part, bonding of the Ardmore basin rocks is by simple adhesion; and, hence, they are easily disaggregated. Gentle crushing and rubbing with a rubber pestle was sufficient to break down all of the rocks without chemical cements, and was the only method used on these samples. Carbonate cement, when present, was dissolved in warm dilute hydrochloric acid; this procedure, of course, also destroys clastic carbonate grains. Silica cement, though present in many samples, caused no difficulty as it occurs as very small overgrowths which are easily broken without breakage of the grains.

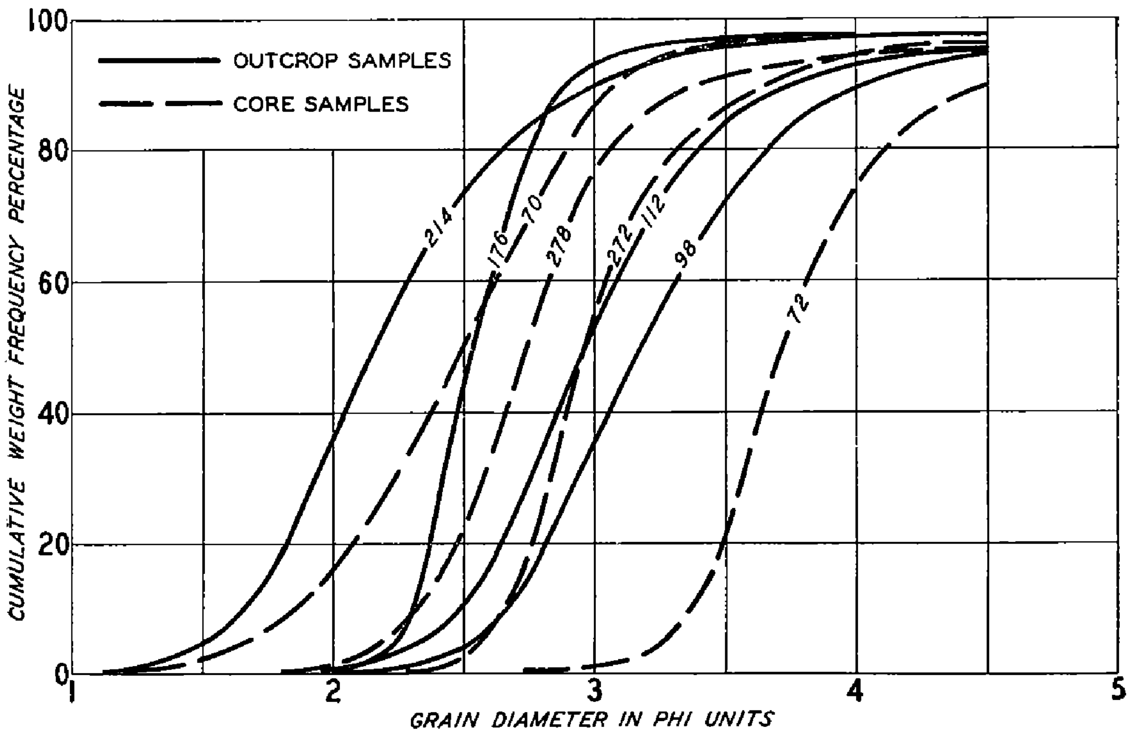


FIGURE 37. Cumulative size frequency distributions of Deese sandstone samples.

All sieved size fractions were checked with a binocular microscope for aggregate grains. Where the proportion of aggregates was estimated to be more than one per cent of the amount sieved, the sample was redone.

Sample size for sieving ranged from 100 to 150 grams; the sieves were U. S. Standard mesh with a fourth root of two, or one-fourth phi, scale; and shaking was on a Ro-tap machine for a period of 15 minutes. The weights of the sample and all sieve fractions were determined to the nearest one-tenth gram.

The phi notation of Krumbein (1934, b) has been used throughout the description of results.

Description of results.—The raw data and summary statistics obtained from the sieving of 56 samples are presented in Table 11, and the size distributions of a representative group of these samples are shown as cumulative curves in figures 36, 37, and 38. The range in grain size is small, and the size distributions are nearly all very similar. The cumulative curves are marked by general smoothness and symmetry, and by extreme steepness of the central part of the curve.

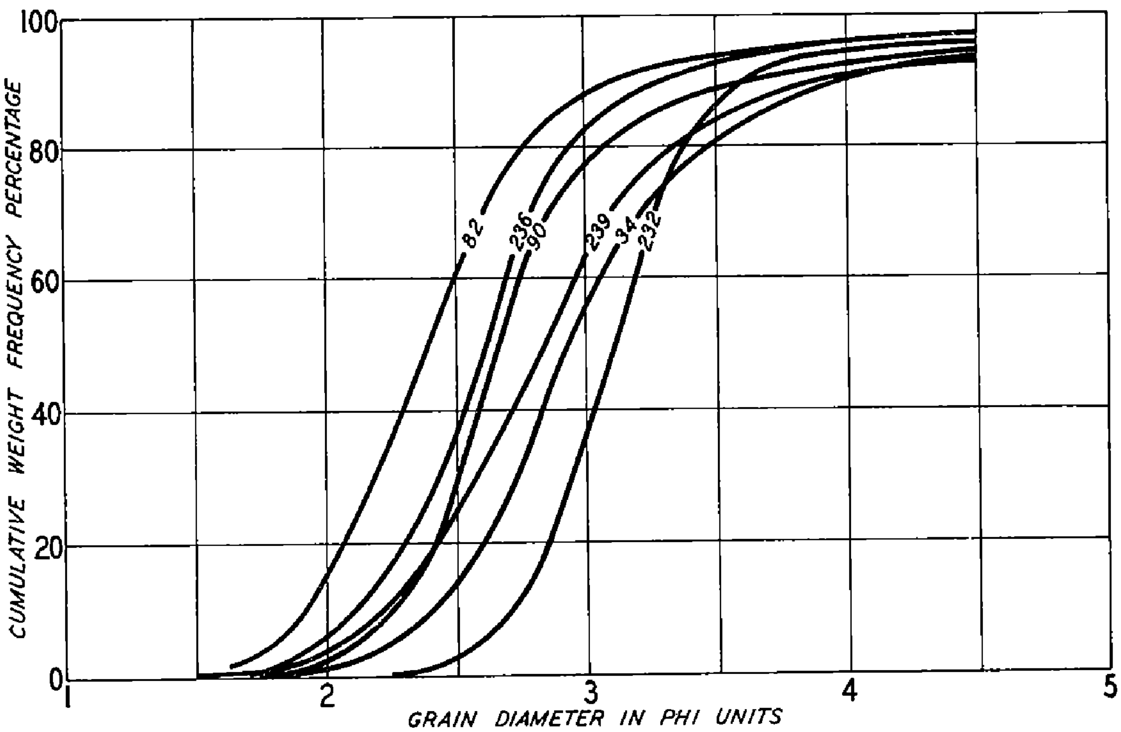


FIGURE 38. Cumulative size frequency distributions of Hoxbar sandstone samples.

However, the distributions are not log-normal. This is best shown by plots of the distribution on probability paper, as in figure 39. Without exception, the samples have the same general pattern. The major portion of the distribution plots as a straight line, indicating approximate log normality; but between the 70th and 98th percentile the slope breaks sharply to the right, indicating an excess of material in the fine grade sizes. Such a distribution has been frequently observed (Pettijohn, 1949, p. 36), and several possible explanations for it are apparent: (1) the sample may be composite; (2) the phenomenon may be due to inefficiency of sieving below about 100 microns; (3) the fine distribution may represent suspension load and the coarse traction load; (4) the fine material may have been trapped in the interstices of the coarser after the latter had been deposited; and (5) the fine

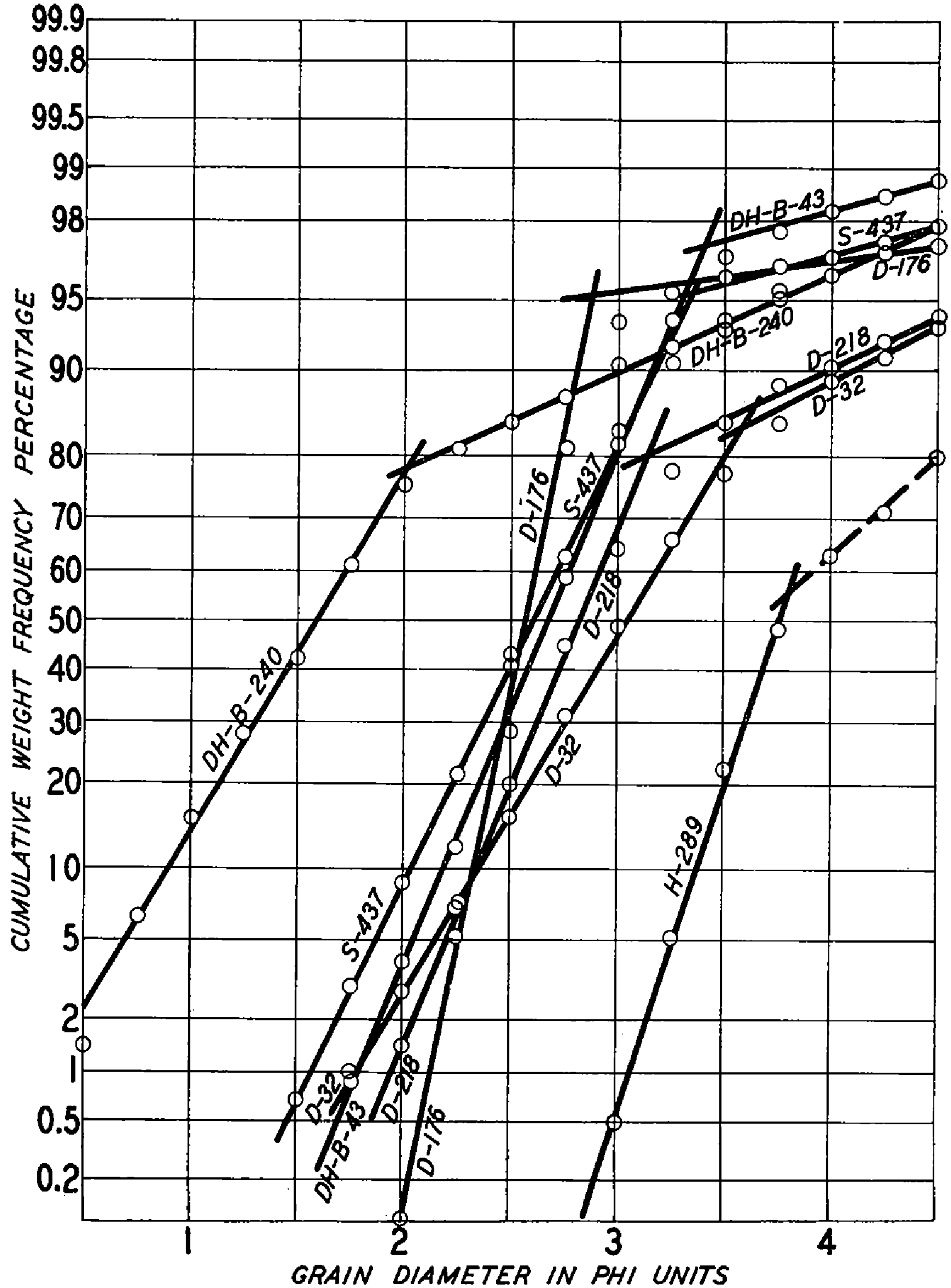


FIGURE 39. Cumulative size frequency distributions of Ardmore sandstones plotted on probability paper.

material may have been sedimented as coarser particles than those which were sieved, *i.e.*, dispersion has reduced the depositional size.

For these samples, it is most likely that the excess of fine material is largely due to breakage of friable rock fragments during disaggregation. The average percentage of material in the pan fraction (<44 microns) is slightly more than five per cent, and the range is 1.1 to 22.4; yet examination of thin sections and thin section size analyses fail to disclose nearly this proportion of such fine grains. In many of the Deese and Dornick Hills samples no grains with a diameter of less than 40 microns can be seen, but the sieve analyses show a pan fraction of five to seven per cent. All of these rocks have ten to fifteen per cent of rock fragments. This suggests that during disaggregation and sieving about a third to a half of the rock fragments are broken sufficiently to pass the finest sieve. Some samples have a matrix of fine-grained minerals, and this is the major contribution to the pan fraction in those samples; but for all samples no correlation is found between the amount of visible matrix and the percentage of silt and clay found in the pan.

Even more definite evidence that the excess of fine material is due to breakage of rock fragments is shown by the high degree of correlation between the percentile at which the distribution breaks from normality and the percentage of rock fragments (determined by the point count method of Chayes [1949]). Figure 40 is a graph of the relationship between the two. The correlation coefficient, "r", (Youden, 1951) of the two variables is -0.91; and the coefficient of determinations, "r"^{2}, indicates the degree of association of the two variables to be more than 80 per cent.

Krynine (1941) recognized the problem of rock fragment breakage in his study of the Bradford sand, and suggested that as a correction a certain proportion of the pan fraction be transferred to the size fraction just coarser than the median. Such an arbitrary correction is not applicable here, however, since the proportion of the pan fraction which is due to crushing of fragments varies from sample to sample.

The most simple correction would be to assume that the size distributions are actually log-normal. Many sediments closely

TABLE 11

Sieve analysis data

Mm.	0.59	0.50	0.42	0.35	0.30	0.25	0.21	0.18	0.15	0.125
Phi.	0.75*	1.0	1.25	1.5	1.75	2.0	2.25	2.5	2.75	3.0
										Springer
S-RC-45						0.2*	1.5	2.9	11.8	26.1
S-165							0.1	0.9	4.1	22.9
S-LA-174									0.1	2.6
S-O-178							0.2	1.7	6.2	16.1
S-LA-190								0.1	0.2	2.4
S-0-195								0.1	0.7	13.1
S-RC-196							0.2	0.9	4.3	22.4
G-198								0.2	1.8	23.1
S-O-242							0.4	3.2	8.4	17.3
										Springer
S-S-78							0.3	3.7	18.0	30.2
S-S-104						2.3	11.0	18.0	18.6	22.0
S-C-402							0.1	0.1	0.6	3.6
S-S-424					2.5	9.8	16.7	14.1	11.3	10.7
S-G-429							0.1	0.2	4.6	17.6
S-S-433								0.2	0.6	7.8
S-S-437				0.8	2.2	6.1	12.5	19.0	22.6	20.1
S-444								0.2	1.9	23.5
S-S-457						1.3	6.0	21.8	35.3	23.1
										Dornick Hills
DH-36							0.2	3.5	18.8	41.1
DH-B-43					1.0	3.0	8.0	16.4	30.6	22.8
DH-140							0.1	1.3	9.6	33.6
DH-186							0.1	1.8	10.6	20.5
DH-220							0.2	2.1	7.1	25.1
DH-221							0.1	0.4	5.0	13.8
DH-222						0.3	0.5	1.4	5.7	22.0
DH-B-240	6.2	9.4	12.3	14.0	18.8	14.8	5.4	3.4	3.1	3.3
										Deese
D-32					0.9	2.0	4.6	8.2	15.8	17.7
D-N-98						0.4	0.8	3.6	12.2	19.1

*Frequencies are given as percentages.

0.105	0.088	0.074	0.0625	0.0526	0.044				
3.25	3.5	3.75	4.0	4.25	4.5	pan	Mdϕ	PDϕ	sϕ
outcrop									
26.1	13.2	10.6	4.0	0.4	1.0	2.3	3.05	0.52	0.39
33.5	22.1	8.4	3.3	0.6	1.1	2.9	3.18	0.41	0.30
24.6	37.5	17.9	7.1	1.6	1.7	6.9	3.40	0.45	0.24
31.8	27.4	10.5	3.4	0.4	0.6	1.5	3.22	0.42	0.36
20.8	29.9	17.0	8.3	4.1	2.0	15.3	3.47		0.28
27.5	30.4	16.7	6.3	1.3	1.0	2.7	3.33	0.42	0.30
23.7	18.6	10.2	4.7	0.8	0.8	2.5	3.12	0.43	0.40
36.4	21.4	8.8	2.9	0.7	1.0	3.8	3.17	0.41	0.28
23.2	22.6	13.0	6.3	1.6	1.4	2.6	3.22	0.55	0.40
cores									
27.4	10.1	3.2	2.0	0.6	0.9	3.6	2.98	0.45	0.28
10.4	7.1	3.4	1.5	0.6	0.9	4.0	2.75	0.68	0.30
29.3	31.4	17.8	6.7	2.3	2.3	5.6	3.37	0.48	0.32
6.4	8.9	6.4	3.9	0.7	1.5	7.1	2.65	1.02	0.84
37.4	17.6	9.0	4.9	0.9	1.6	6.0	3.14	0.56	0.27
21.1	38.8	17.7	6.9	1.2	1.7	4.0	3.38	0.42	0.29
7.3	3.3	1.6	1.5	0.3	0.6	2.1	2.60	0.60	0.44
34.0	21.7	7.8	3.3	1.2	1.5	4.9	3.18	0.46	0.24
5.3	1.8	0.8	1.0	0.5	0.4	2.7	2.64	0.38	0.30
outcrop									
19.7	6.9	2.6	1.4	0.7	1.2	3.9	2.93	0.43	0.25
11.8	3.2	0.8	0.7	0.3	0.2	1.3	2.66	0.48	0.38
27.9	12.4	5.9	2.7	0.8	1.1	4.7	3.04	0.48	0.26
24.4	16.8	9.4	5.7	1.6	1.6	7.5	3.19	0.70	0.33
22.1	15.6	9.3	5.5	3.6	1.3	8.0	3.17	0.72	0.36
24.4	17.2	12.5	9.6	2.9	3.7	10.5	3.33	0.83	0.34
13.3	9.2	8.6	8.2	3.4	4.9	22.4	3.43		
1.3	1.1	1.9	1.2	0.9	1.0	1.3	1.61	1.07	0.54
outcrop									
17.3	11.2	7.0	4.5	1.9	2.5	6.5	3.00	0.86	0.54
21.1	13.4	13.7	4.6	3.0	2.2	5.3	3.15	0.71	0.43

TABLE 11 (Continued)

Sieve analysis data

Mm.	0.59	0.50	0.42	0.35	0.30	0.25	0.21	0.18	0.15	0.125
Phi.	0.75*	1.0	1.25	1.5	1.75	2.0	2.25	2.5	2.75	3.0
D-112						0.2	3.4	6.4	20.2	22.4
D-122					2.5	7.1	15.7	18.9	23.7	12.0
D-130						0.2	2.3	18.8	43.5	22.7
D-151					0.6	3.3	11.6	20.0	25.1	19.2
D-176						0.1	5.1	38.0	37.3	13.3
D-181						0.2	1.5	10.6	39.2	28.1
D-DK-207						0.2	1.6	5.5	15.1	30.1
D-DK-214			1.2	3.6	10.1	21.6	20.7	16.0	10.3	6.8
D-218					0.1	1.3	5.5	12.9	25.0	19.3
										Deese
D-70			0.5	1.7	4.5	9.3	15.9	17.3	19.1	18.6
D-72							0.1	0.1	0.1	0.4
D-75					0.1	0.6	3.1	6.7	19.6	23.3
D-272							0.1	2.2	18.6	33.8
D-278					0.1	0.8	5.5	17.2	28.0	26.1
D-284					0.8	3.5	7.3	10.3	9.9	12.3
D-285						0.8	5.1	12.3	25.8	25.2
D-286									0.1	4.3
D-292								0.2	1.8	7.3
D-305								0.1	0.5	4.3
										Hoxbar
H-34					0.1	1.0	4.2	8.3	18.2	23.5
H-82					3.6	11.7	20.6	24.2	19.4	9.0
H-84		0.9	1.3	2.2	5.7	10.8	14.6	16.6	13.9	11.0
H-Z-90					0.6	2.0	6.5	21.4	30.5	16.2
H-232							0.5	2.3	8.8	25.8
H-236				1.9	3.1	5.9	10.8	19.2	28.1	14.5
H-239					0.8	3.4	6.9	14.0	18.5	20.5
										Hoxbar
H-289										0.5
H-290								0.1	0.4	4.8

*Frequencies are given as percentages.

0.105	0.088	0.074	0.0625	0.0526	0.044				
3.25	3.5	3.75	4.0	4.25	4.5	pan	Mdϕ	PDφ	sϕ
19.4	11.8	5.5	3.4	1.1	1.1	5.0	2.97	0.64	0.41
7.5	4.4	1.9	1.5	0.3	0.8	3.7	2.53	0.40	0.30
4.2	1.5	1.1	0.9	0.6	0.6	3.4	2.67	0.34	0.22
6.4	3.5	2.2	1.7	0.8	0.7	4.9	2.64	0.70	0.36
1.5	0.8	0.4	0.4	0.1	0.4	2.3	2.54	0.30	0.19
9.9	3.4	1.2	1.1	0.2	0.8	3.7	2.74	0.40	0.26
19.5	10.1	5.7	3.6	1.1	1.1	6.4	2.98	0.66	0.35
3.1	1.8	1.2	0.7	0.3	0.5	2.0	2.15	0.68	0.44
13.6	6.7	4.3	1.7	2.2	1.5	6.0	2.81	0.76	0.40
cores									
6.6	2.2	0.8	1.0	0.2	0.4	2.0	2.52	0.62	0.50
3.7	18.3	31.3	31.2	6.8	8.1	9.9	3.72	0.67	0.30
20.8	11.4	5.2	3.7	1.4	2.5	1.7	2.97	0.60	0.42
22.1	9.2	5.6	2.3	1.4	1.1	3.6	2.97	0.52	0.25
9.4	3.4	1.9	1.5	0.7	1.0	4.3	2.74	0.55	0.32
13.5	15.1	9.5	6.4	1.9	2.7	6.8	3.12	0.95	0.74
12.4	5.0	2.7	1.9	0.8	1.1	7.0	2.81	0.74	0.37
13.2	22.2	25.9	9.0	4.4	6.1	14.9	3.58	0.86	0.32
11.4	13.8	18.1	22.9	....	11.3	13.3	3.72	0.91	0.60
31.5	28.0	15.3	4.7	2.7	4.3	8.4	3.36	0.78	0.29
outcrop									
15.9	7.2	7.5	4.8	0.4	1.7	7.1	2.93	0.78	0.32
3.5	1.9	0.8	1.1	0.4	0.8	2.9	2.40	0.58	0.36
5.9	4.7	3.7	2.8	1.0	1.6	3.3	2.47	0.95	0.60
7.8	3.6	1.9	2.2	0.7	1.2	5.4	2.66	0.70	0.42
29.4	19.4	6.2	2.0	0.7	0.8	4.2	3.11	0.44	0.36
6.5	2.6	2.1	1.3	0.6	0.7	2.8	2.58	0.64	0.41
13.1	6.8	4.3	2.9	1.1	1.7	6.1	2.83	0.84	0.48
cores									
4.7	16.8	25.5	14.9	8.7	7.8	21.1	3.76	....	0.30
23.1	26.6	15.1	7.2	4.1	5.4	13.2	3.45	....	0.28

approach this, particularly those which are texturally "mature" (Krumbein, 1938). The fact that the thin section analyses approach normality closely and the high degree of correlation between the percentage of rock fragments and the percentile at which the sieved distributions depart from normality strongly suggest that the size distribution in these samples is not significantly different from a normal one. If we further assume that the broken rock fragments have the same size distribution as the balance of the sample, the best approach to the true distribution is given by continuing the probability plot as a straight line beyond the point of inflection. This has been done for all samples, and the values for the standard deviation given on Table 11 have been read off such plots on the basis of the assumption of normality. Further, with these assumptions the median value becomes the same as the mean value. The percentile deviation values of Table 11 are based on the uncorrected data.

For individual samples the range in medians by sieving is 1.61 to 3.76 phi. However, most of the medians fall within a very narrow range. Of the 56 samples 48 have a median between 2.5 and 3.5 phi. Only one sample would be classed as a medium sand (1 to 2 phi), 27 are fine-grained (2 to 3 phi), and 28 are very fine-grained (3 to 4 phi).

Size sorting of the individual samples is excellent; for nearly one-half of the samples 90 per cent of the grains fall within one Wentworth size grade. In Table 11 two "sorting coefficients," the phi percentile deviation (PH^{ϕ}) and the phi standard deviation (s^{ϕ}), are given for each sample. The phi percentile deviation is the statistic used by Griffiths (1951, 1952) in his study of size relationships in a large number of samples, and is thus convenient for comparative purposes. It is a measure of half the spread between the ninetieth and tenth percentile diameter values as read off the cumulative curve. The values given are not corrected for any technique errors.

The standard deviations were determined graphically from plots of the size distributions on arithmetic probability paper (figure 39) as suggested by Otto (1939). Graphical determination of the standard deviation requires the assumption that the distribution is normal; and, hence, the standard deviation is a measure of sorting which has been corrected for the skewness brought about by the breakage of rock fragments. It should be a better basis for comparison of samples than the percentile deviation since it corrects an error in technique which may have considerable effect on the percentile deviation, particularly in determination of the ninetieth percentile.

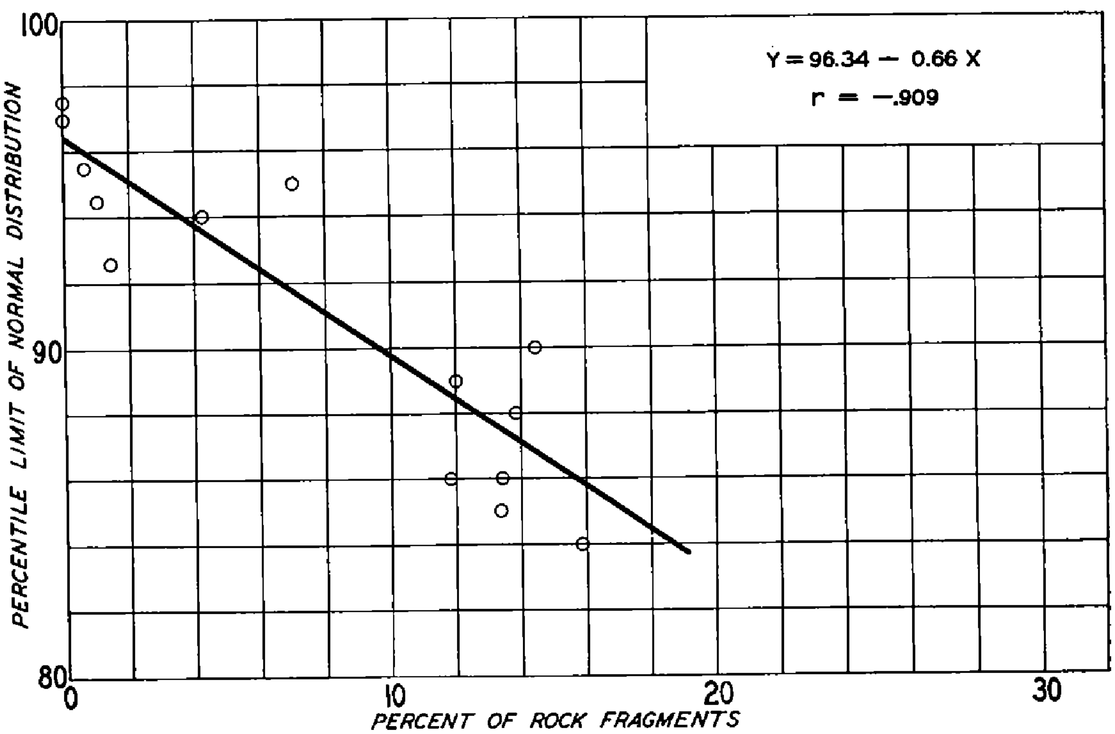


FIGURE 40. Relationship between proportion of rock fragments and nature of cumulative curves for Ardmore sandstones.

All of the Ardmore basin sandstones fall within Griffith's (1952) best sorted sands. Only two have percentile deviation greater than one phi, and many are less than one-half phi. They are, thus, comparable in sorting to the best sorted orthoquartzites.

Thin section size analysis.

Technique.—Size distributions were determined from thin sections by measurement with a micrometer ocular. Measurement was of the apparent long axes of randomly selected grains. Randomness was obtained by measuring all grains intersected by short lines drawn on the cover glass and spaced uniformly over the area of the thin section.

For most samples 100 grains were measured; for 1 sample, 120 grains; for 1, 150 grains; and for 2, 200 grains. The precision obtained by any given number of measurements is dependent

TABLE 12

Thin section size analysis data.

mm.	0.59	0.50	0.42	0.35	0.30	0.25	0.21	0.18	0.15	0.125
Phi	0.75*	1.00	1.25	1.50	1.75	2.00	2.25	2.50	2.75	3.00
S-RC-45					1*		5	14	24	19
S-LA-48								1	2	6
S-P-49										
S-S-77	2	5	9	16	20	15	13	9	2	6
S-RC-175										
S-O-178							5	3	20	23
S-402						1	1	1	8	22
S-S-437				4	9	8	24	24	12	8
S-S-455						1	2	8	12	32
S-S-457				1	4	14	17	17	16	10
S-S-459					3.3	8.3	20.1	16.6	17.5	10.0
S-S-462							3	5	16	27
S-S-464									1	5
DH-36						1	12	15	25	16
DH-B-43					3	7	16	27	21	13
D-32					2	6	7	11	14	15
D-DK-40	2.0	3.0	5.5	23.5	26.0	17.5	11.0	6.5	2.0	1.5
D-284			0.7	2.7	4.0	10.0	14.7	6.7	12.6	11.3
H-84		1.0	2	3	9	11	16	12	11	8

*Frequencies are given as percentages.

upon the variance (sorting) of the sample; so the precision is not uniform for the several samples. For the samples in Table 12 the 95 per cent fiducial interval of the mean ranges from ± 0.07 phi (D-DK-40) to ± 0.13 phi (D-284).

Description of results.—Table 12 presents the data obtained from nineteen samples, and figure 41 shows cumulative curves for five representative size distributions.

The range in means is from 1.79 phi to 4.16 phi, which is slightly greater than that obtained for the sieved samples. This arises from the inclusion of several samples which were thought too fine-grained for sieving. However, as for the sieved samples,

0.105	0.088	0.074	0.0625	0.053	0.044	0.037	0.031				
3.25	3.50	3.75	4.00	4.25	4.50	4.75	5.0	>5	$\overline{X}$	s	n
13	12	6	2	1	2	1			2.84	.52	100
11	20	23	13	9	7	5	1	2	3.67	.54	100
5	14	20	27	16	16	7	3	8	4.16	.47	100
3									1.79	.57	100
2	8	18	21	20	17	7	2	5	4.05	.46	100
19	13	7	4	3	2	1			3.06	.51	100
19	17	11	5	5	4	2	1	3	3.32	.60	100
4	3	1	2	1		1			2.35	.54	100
18	10	7	6	3					3.02	.45	100
8	7	2		2		1		1	2.53	.59	100
10.9	9.1	2.5	0.9	0.8					2.72	.53	120
15	13	8	4	4	2	1	1	1	3.13	.57	100
10	14	14	19	12	9	9	4	3	3.85	.58	100
15	9	5	1	1					2.78	.46	100
6	5		2						2.52	.44	100
14	12	11	4	2	1	1			2.92	.62	100
1.0	0.5								1.70	.47	200
9.3	8.7	9.3	4.7	0.7	2.7	1.3	0.7		2.77	.78	150
13	5	1	5	2		1			2.49	.70	200

nearly all sandstones were either fine- or very fine-grained.

The size distributions as shown by the cumulative curves are very much like one another, but they differ from the sieved distributions, particularly in the greater slope of the central part of the curves. This difference is due, however, to the techniques used and does not reflect any fundamental difference in the character of the sandstones.

Comparison of sieve and thin section size distributions.—Because of the effects of sectioning of grains and of other factors (Rosenfeld, Jacobsen, and Ferm, 1953) the results of the two measurement techniques are not directly comparable, and to make them comparable some sort of conversion is necessary. This problem has given rise to a considerable literature (Krumbein, 1935; Greenman, 1950, 1951; Chayes, 1950; Pelto, 1952), and has recently been treated in detail by the writer and others (*op. cit.*).

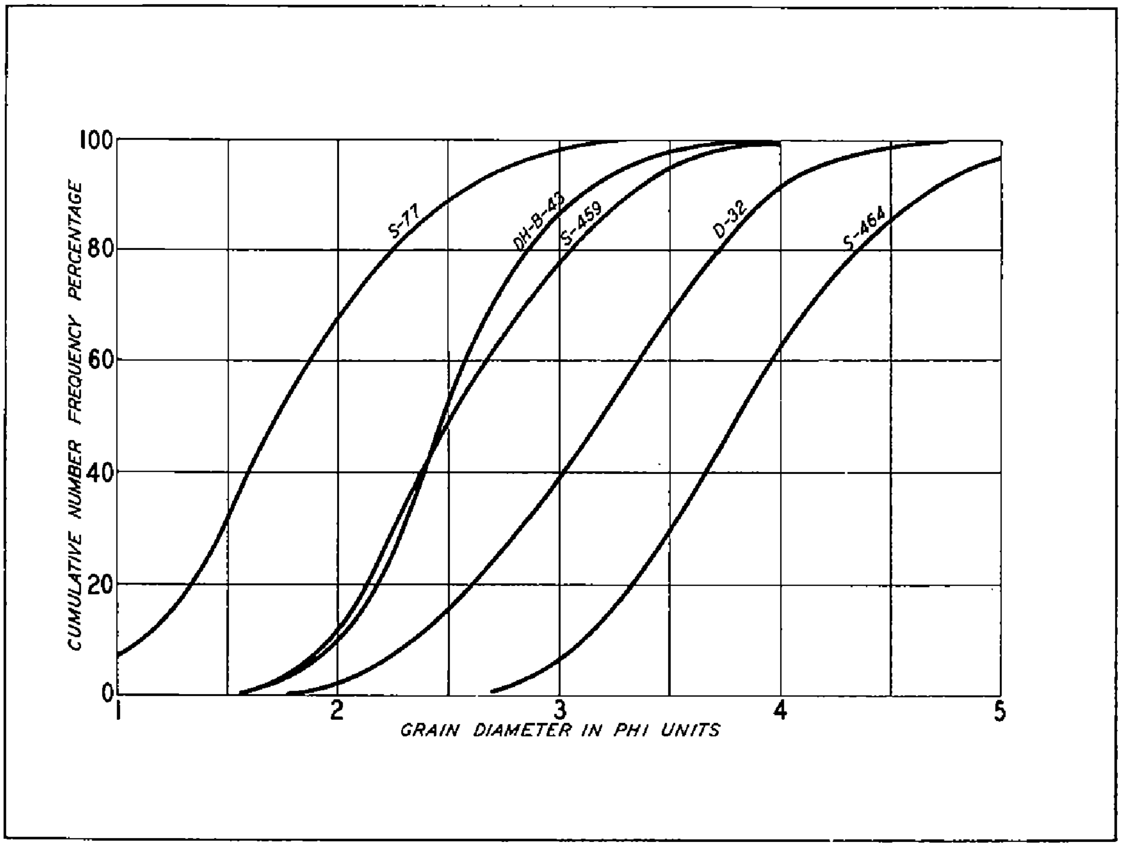


FIGURE 41. Cumulative size frequency distributions determined by thin-section measurements.

In this latter study evidence was presented that no known combination of theoretical corrections for the effect of sectioning, for the use of number instead of weight frequency, or for sieve calibration could bring the two techniques into agreement. It was concluded there was no unique factor which could be applied to a technique, but instead the necessary conversion varied greatly with different operators and rock types; the only satisfactory conversion was one derived by empirical comparison of the results of the two techniques within the experiment for which the conversion was to be used.

Accordingly, both sieve and thin section analyses were made on ten samples. The results of the two determinations expressed as means and standard deviations[1] are presented in Table 13. Krumbein's (1934) phi notation is used.

[1] The standard deviations for the sieve analyses were determined graphically from a plot of the size distribution on arithmetic probability paper.

For all samples the thin section analyses show a coarser mean size and a lesser degree of sorting than do the sieve analyses. The difference in mean size is due to a complex of interacting factors (Rosenfeld, *et al, op. cit.*), and perhaps the difference in standard deviation may be similarly accounted for; but it is probable that the increase is mainly due to the effect of sectioning. As the plane of the section has no necessary relationship to the maximum diameter of the grains, even uniform grains would show a size distribution if cut at random, as are the grains in a thin section.

TABLE 13
COMPARISON OF SIEVE AND THIN SECTION SIZE ANALYSES

Sample	Sieve analyses		Thin section analyses	
	Mean	Standard deviation1	Mean	Standard deviation
H-84	2.49	0.60	2.47	0.70
S-437	2.60	0.44	2.35	0.54
S-457	2.64	0.30	2.53	0.59
DH-S-43	2.66	0.38	2.52	0.44
DH-36	2.93	0.25	2.78	0.46
D-32	3.00	0.54	2.92	0.62
S-RC-45	3.05	0.39	2.84	0.52
D-284	3.12	0.74	2.77	0.78
S-O-178	3.22	0.36	3.06	0.51
S-402	3.37	0.32	3.32	0.60

The relationship between the two sets of statistics appears to be linear and can best be expressed as the equation of the line of best fit. This was computed by the method of least squares (Youden, 1951). Figure 43 is the plot of mean diameters of sieving *vs.* thin section (sieve median is considered the independent variable) and includes the computed line.

The correlation coefficient "r" for the means is 0.9381, and the equation of the line is $y = -0.0112 + 0.9515\ x$, in which "y" is the mean as determined in thin section, and "x" is the mean from sieve analysis. However, the intercept value, -0.0112, is not significantly different from 0.0; and the regression coefficient, 0.9515, is not significantly different from 1.0 ($t = 0.3757$, 8 D.F., Youden, 1951). Hence, the average difference between samples of 0.15 phi is as satisfactory a conversion factor as the regression line. To compare thin section means to sieve means it is only necessary to add 0.15 to the thin section value.

The correlation coefficient for the standard deviations is 0.8189, and the equation of the line is $y = 0.3285 + 0.5727\ x$, where "y" is the thin section value, and "x" is the sieve value. This regression line is the best available conversion factor for standard deviations of the two techniques. The standard error of estimate (Youden, 1951) is 0.065.

As an illustration of these conversions, consider a sample such as S-104 which has a mean of 2.75 phi and a standard deviation of 0.30 phi as determined by sieving; then to convert to thin section units, the mean is obtained by subtracting 0.15 from 2.75, giving a corrected mean size of 2.60 phi. The standard deviation is obtained by substituting 0.30 for "x" in the second equation above. Thus, the corrected standard deviation equals $0.3285 + 0.5727 (0.30)$, or 0.49.

Since there is no basis for considering the results of one technique more fundamentally correct than that of the other, the correction equations may be used with either "x" or "y" as the unknown; and the conversion may be made either from sieve to thin section, or from thin section to sieve data.

Combined sieve and thin section results.—The best available estimate of grain size distribution in the Ardmore basin sandstones is the distribution obtained by combining the results of the 67 individual samples. This combined or total distribution is shown as a unique frequency curve (Brotherhood and Griffiths, 1947) in figure 44.

In computing this curve the thin section size analyses were converted to make them comparable to the sieve analyses, as already described; and all sieve analyses were corrected for the effect of rock fragment breakage by assuming a log normal size frequency distribution.

The mean grain size of the Ardmore sandstones as estimated from 65 samples is 3.04 phi, and the standard deviation is 0.62 phi. The total distribution is very nearly a normal one as can be seen in figure 44, the skewness is only 0.09; and the kurtosis is 1.04. Because of the large amount of data, even this small departure from normality is statistically significant (chi square 184.70, D. F. 22, $P < 0.999$).

With the very narrow range of grain size in the basin, the differences between formations must necessarily be small; and, if present at all, are likely only to be detected by examining a large number of samples.

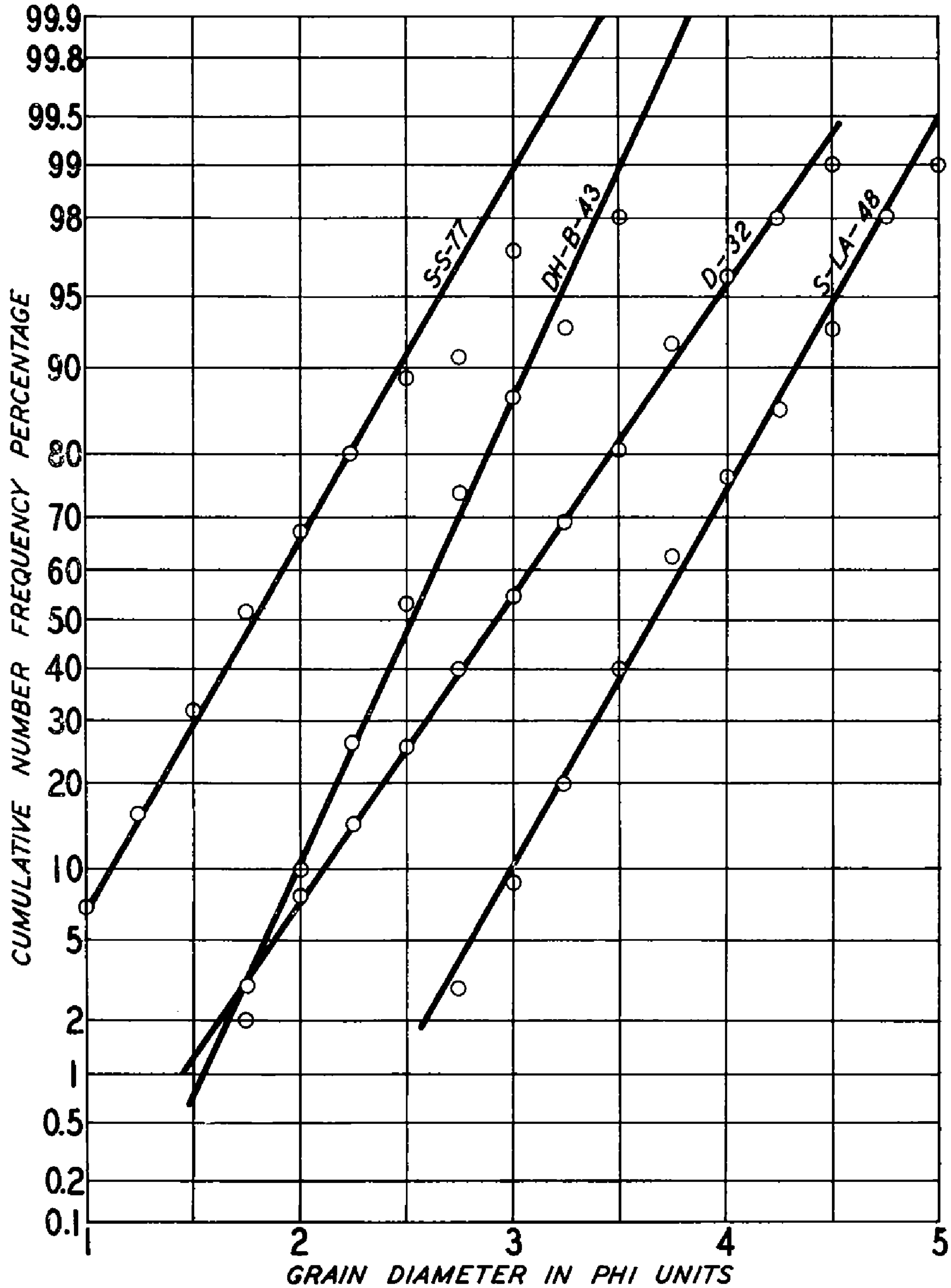


FIGURE 42. Cumulative size frequency distributions determined by thin-section size analyses and plotted on probability paper.

Statistics of the sample means grouped by formational units are tabulated in Table 14.

The formational means in the table are the arithmetic average of the several sample means, and the standard error of the mean is the conventional measure of variation among samples.

The differences between units are of the same order or even smaller than the standard error, and it is not obvious that any of these differences are significant. However, the differences may be examined by means of the "t" test (Youden, 1951); and the probability determined that variations of the observed magnitude could be due to chance alone. For this application the "t" test is essentially a comparison of the observed differences with the precision of the difference.

TABLE 14
MEAN GRAIN SIZE OF SANDSTONE SAMPLES

Formation	Mean	Number of samples	Standard error
Springer	3.20	0.458	26
Springer outcrop	3.47	0.424	12
Springer cores	2.99	0.485	14
Dornick Hills (outcrop only)	2.92	0.580	8
Deese	2.91	0.397	22
Deese outcrop	2.70	0.363	13
Deese cores	3.15	0.425	10
Hoxbar (outcrop only)	2.71	0.260	7

The Springer taken as a whole is finer-grained than any of the formations (compared with Deese, $t = 2.185$; 46 D.F., $P < 0.05$), but there is no significant difference between any two of the Dornick Hills, Deese, and Hoxbar formations, at least with the present number of samples. Furthermore, the Dornick Hills, Deese, and Hoxbar samples are not significantly different in mean size from the Springer core samples.

Observed differences between surface and subsurface samples within formations are as great as differences between formations. For example, the average median of the Springer surface samples is 3.46 phi; and of the Springer cores, 2.99 phi. The probability of such a difference arising by chance is less than five per cent ($t = 2.56$, 22 D.F.); and, hence, it can reasonably be concluded that the difference is real. It is possible the difference could be due to sampling. However, Lucas (1934) has reported 153 size analyses of surface Springer samples; and he found no samples as coarse as many cores.[1] With the supporting evidence of Lucas' samples it seem highly probable that Springer sandstones from the oil fields of the western part of the Ardmore basin are, on the average, coarser than the sandstones which crop out around Ardmore.

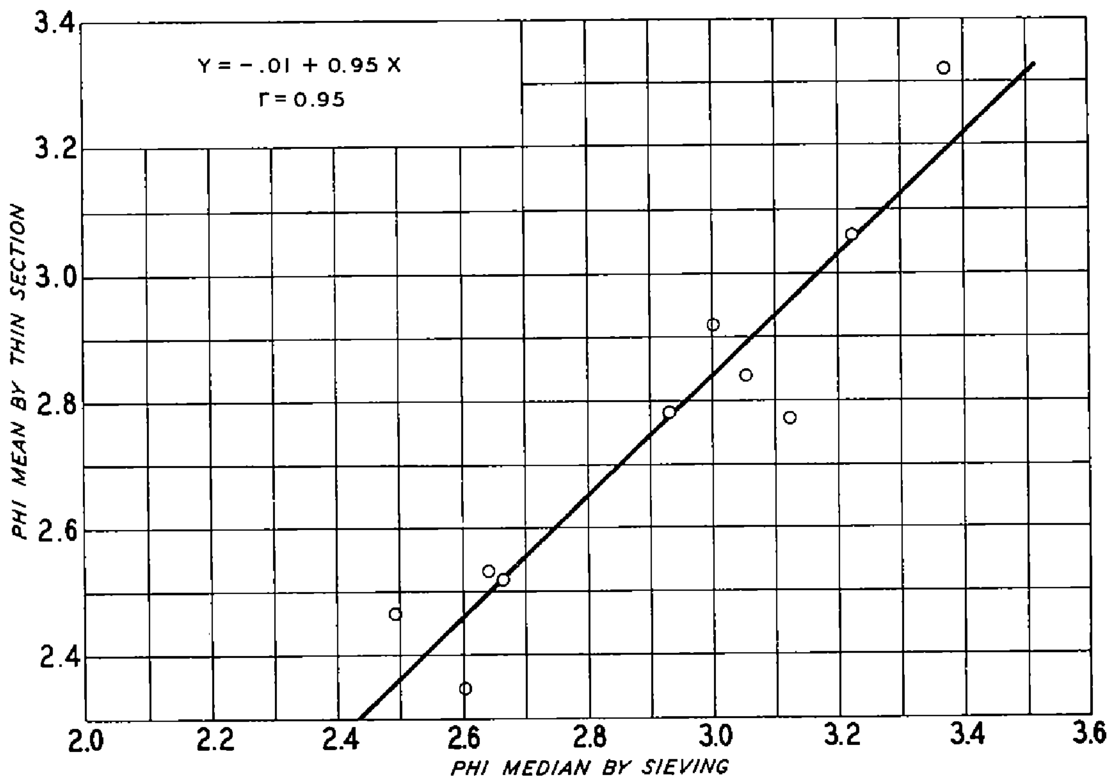


FIGURE 43. Relationship between median grain diameter determined by sieving and mean grain diameter determined by thin-section size analyses.

Similarly, if we test the Deese surface and core samples, we find that the probability of the observed difference being due to chance is less than one per cent ($t = 6.62$, 20 D.F.). However, there is a strong suggestion here that the samples are biased; four of the twelve Deese surface samples are from the Devils Kitchen conglomeratic member, and the sandstones associated with conglomerates are characteristically coarser than average, thus, tending to bias the average toward the coarse side. If only one or two of these Devils Kitchen samples had been included, the difference would not have been significant.

[1] Because of differences in technique Lucas' samples cannot be closely compared to these, but his samples had an average median diameter of between 3.40 and 3.50 phi. The coarsest sample was about 3.0 phi.

No serious attempt was made to compare variation in size within a formation and variation within a single sandstone bed, but some data is available to evaluate this relative variation. Five samples were analyzed from a 200-foot diamond core of the Sims member of the Springer formation. These are 455, 457, 459, 462, and 464; and the data are given in Table 15. The mean of these five samples is 3.05 phi, and the standard error is 0.51. This standard error is of the same magnitude of those for the several formations as given in Table 14; and, thus, it can be seen that the amount of variation within this particular sandstone based on five samples is no less than the variation within the entire formation.

TABLE 15
GRAIN SIZE STATISTICS FOR FIVE SAMPLES FROM A CORE OF THE SIMS MEMBER OF THE SPRINGER FORMATION

Sample Number	Depth	Mean	Standard deviation
455	4973	3.02	0.45
457	5012	2.53	0.59
459	5049	2.72	0.53
462	5122	3.13	0.57
464	5139	3.85	0.58

Conclusions.—Grain size analyses have been of little use either in differentiating stratigraphic units or in defining rock types. What value the size distribution has in defining the rock types is almost entirely a reflection of the dependency of size upon mineral composition. For example, the rocks which have an appreciable content of rock fragments also have a distinctive size distribution, but only because the rock fragments were broken during size analyses. In general, the differences which are apparent in texture are even more apparent in mineral composition and much more readily studied.

Probably the most significant fact concerning grain size is that the Springer reservoir sandstones are coarser-grained than the remainder of the Springer sandstones. As will be seen, this is only one of several ways in which these rocks are distinctive, and can be used as a partial definition of a reservoir facies.

Whether there is any special significance to the near normality of the total grain size distribution is not known. It may be a negative sort of evidence as to the source area. If the distribution were bimodal or strongly skewed, a complex or multiple source area would be indicated; but the normal distribution cannot be interpreted to indicate a single simple provenance.

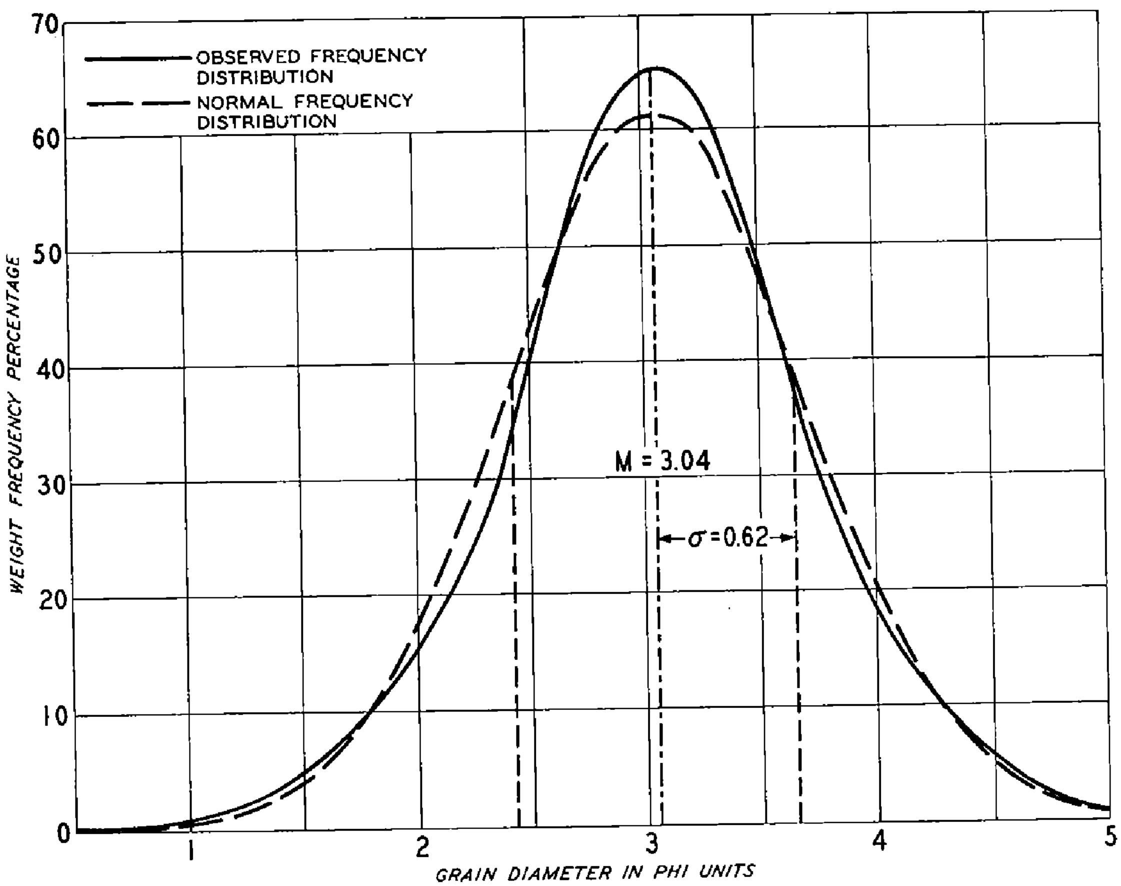


FIGURE 44. Total grain size frequency distribution for 67 samples.

SEDIMENTATION

Rate of Deposition

The rate of deposition is important in sedimentation in so far as it controls the time available for all processes modifying the detritus in the area of deposition. These, along with provenance, are the fundamental factors in sedimentation.

Necessarily, any estimate of the rate of deposition of ancient sediments is subject to major absolute errors. Nevertheless, Kay (1951) has given a basis for comparing rates of sedimentation in his estimate of the rates of subsidence in a number of geosynclinal areas. By using the same radioactive time scale and by using maximum thickness as did Kay, a rate of sedimentation can be computed which should be qualitatively comparable to those used by Kay.

Table 16 gives the maximum known thickness of the sediments in the Ardmore basin and duration of the Pennsylvanian epochs as given by Kay.

As a correlation with the epochs used by Kay, the lower Pennsylvanian sequence is considered to include the Goddard and Springer formations; the middle Pennsylvanian sequence from the top of the Springer to the Arnold limestone member of the Deese formation; and the upper Pennsylvanian, the remainder of the system.

TABLE 16
THICKNESS OF SEDIMENTS IN THE ARDMORE BASIN AND DURATION OF PENNSYLVANIAN EPOCHS

Epoch	Duration (millions of years)	Thickness of sediments (feet)
Upper	10	7,500
Middle	15	7,500
Lower	15	5,000
	40	20,000

Subject to errors in the estimates of stratigraphic thicknesses and the duration of Pennsylvanian time, the overall rate of sedimentation is 500 feet per million years for a total of 40 million years. The rate is not uniform but increases with time. For the lower Pennsylvanian formations the average rate is 333 feet per million years; for the middle Pennsylvanian the average rate is 500 feet per million years; and for the upper Pennsylvanian, 750 feet per million years.

These are especially rapid rates considering the duration of time for which they extend. For comparison, in the Ordovician eugeosyncline of the Appalachian area the rate was less than 350 feet per million years. In the Colorado late Paleozoic "zeugeosyncline" the rate was 200 feet per million years. In the California Miocene-Pliocene "epi-eugeosyncline" the rate was 600 feet per million years; in the Pottsville miogeosyncline of West Virginia the rate was 200 feet per million years; and in the Triassic so-called "taphrogeosyncline" of New Jersey the rate was 1,500 feet per million years. Only in the Triassic fault blocks are the accumulation rates appreciably greater for comparable durations. If these comparisons are valid, Ardmore sedimentation was ultrarapid; and it is, thus, likely that the detritus underwent minimum modification in the process of sedimentation. The change from the orthoquartzitic sediments of the Hoxbar to the arkoses of the overlying Pontotoc probably does not reflect any fundamental change in the processes of sedimentation but only the exposing in the source area of large areas of granite.

SEDIMENTARY PROCESSES

The Conglomerates

The presence of unstable constituents such as limestone fragments, the strongly bimodal size distribution, and the localization of the conglomerates along the basin margin are considered the significant features of the conglomerates. In all of these features the conglomeratic units resemble fanglomerates, and it is believed that the sedimentary processes which formed them were essentially analogous to those which produce a modern day fanglomerate.

The conditions required to produce a rock with the characteristics of the Ardmore basin conglomerates must necessarily include (Krynine, 1950, p. 154): (1) erosion which directly and powerfully attacks bedrock rather than only mantle; (2) transportation which is capable of moving coarse detritus without sorting it; and (3) deposition which is abrupt and localized.

All of these are chiefly dependent upon local topography, and can be expected in any area of steep relief and abrupt changes in slope. That the relief need not be high is shown by the fact that the small streams of the present Arbuckle Mountains, where the local relief is mostly under 200 feet, carry detritus which is nearly indistinguishable from that of the Pennsylvanian conglomerates. These streams for most of the year are small and carry almost no sediment; however, for short periods during the spring they become powerful torrents which transport sand, pebbles, and cobbles as a unit. These flash floods dissipate as rapidly as they form, leaving behind gravelly detritus which is texturally similar to that of the conglomerates. If there were continued intermittent uplift of the Arbuckles along a well-defined line such as a fault, these streams would leave a deposit with similar mineral composition and texture to the Bostwick conglomerate.

An intermittently active fault scarp is the ideal environment for the development of fans, inasmuch as it produces both a perfect localizing factor and the necessary relief. Probably all of the Ardmore conglomerates originated from structurally controlled topography, mostly basin margin fault scarps.

Wherever the basin margin is known to be faulted, conglomerates are present at one or more levels. The Jolliff and Bostwick conglomerates occur along the faulted east margin of the Criner Hills, and subsurface conglomerates are known all along the faulted south edge of the western part of the basin. On the other hand, on the south flank of the Arbuckle Anticline, where there is only minor faulting, no conglomerates are known. This does not imply lack of uplift, but only lack of the sharply localized uplift necessary to yield the conglomeratic fans. In the northern part of the Arbuckle Mountains (Hunton Anticline) where faulting was going on during Pennsylvanian sedimentation, similar fans were deposited.

The prominent Devils Kitchen and Rocky Point conglomerates cannot be attributed to any single known fault. They both show a component of coarsening to the southeast, but their three-dimensional relationships are not well enough known to recognize their source. They are wedge-shaped deposits texturally and mineralogically similar to the other conglomerates, and there seems to be no necessity to assign them to any greatly different mode of deposition or provenance.

The emphasis on the importance of faulting in the formation of the conglomerates does not, of course, imply that faulting was of the same relative importance in the deformation of the basin margins. The conglomerates make up only a few per cent of the basin rock, and the balance of the sediments presumably represent deposition from less sharply localized uplifts.

Because of erosion along the uplifted margins of the basin or later deposition, none of the conglomerates can be seen in direct contact with the fault block from which they originated; but the Bostwick and Jolliff conglomerates do occur within a few hundred yards of their source area. However, on the north flank of the Arbuckle Mountains a fault plane with a similar conglomerate is well exposed along a series of high road cuts on U. S. Highway No. 77 (Collings Ranch conglomerate, Ham, 1954).

The interbedding and lateral interfingering of the conglomerates with normal marine sediments plus the moderately well-developed bedding suggest that the conglomerates were probably deposited in shallow marine waters. It is even possible that they are in part the product of direct wave erosion of sea cliffs. However, the mixed nature of the detritus and the considerable lateral changes in thickness and coarseness are more in keeping with erosion and transportation by a number of short high-gradient streams. Under either condition it is likely that the shoreline closely coincided with the margin of the uplift during the deposition of the conglomerates.

Destruction of the less durable portions of the detritus took place during transportation, and it follows that there is a progressive concentration of the more resistant elements in a basinward direction. Thus, as the conglomerates are observed outward from the basin margins, shale pebbles are the first to disappear, then limestone fragments, and finally chert. The outer edges of all the fans consist almost entirely of very pure quartz sand. It should be recognized, however, that the distance over which some of the conglomerates change character (the Bostwick, for example) was originally appreciably greater than at present as tight folding has brought initially widely separated parts of the fan into close juxtaposition.

The determining factor in the formation of the Ardmore conglomerates was the sharp uplift of the basin margins during sedimentation. Once the structural mechanism was in operation all of the features of the conglomerates followed automatically.

The significance of the conglomerates lies in the fact that they show deformation was concomitant with sedimentation, and each records a local episode in the orogeny of the region. Inasmuch as the conglomerates occur throughout the greater part of the Pennsylvanian stratigraphic section, it follows that the diastrophism was not concentrated in one or two short periods but was spasmodic throughout a long period of Pennsylvanian sedimentation. It is these marginal accumulations which record the orogenic history of the region, and, as pointed out by Gilluly (1949), not the minor unconformities within the basin.

The Sandstones

Deposition of the Ardmore sediments seems to have been largely, but not entirely, in a marine environment. Fossils occur throughout the section; in the Springer formation there is an abundant microfauna though larger forms are common only very locally; in the younger formations there is both an abundant micro- and macrofauna. Direct evidence of some non-marine sedimentation is found in two small occurrences of coal, one in the Dornick Hills formation of the eastern part of the basin and the other in the Hoxbar formation at the south edge of Ardmore. It is possible, even probable, that other local areas of subaerial deposition are present; but there are no satisfactory criteria for identifying them.

The Springer basin of deposition was for the most part a restricted somewhat stagnant sea. Numerous features suggest a low level of oxidation in the area of accumulation. Among these are the prevailing dark gray color of the sediments, the abundant carbonaceous material, and the common occurrence of siderite and pyrite in the shales and sandstones.

Massive unwinnowed, and thin-bedded shaly sands make up the bulk of the Springer sandstones; and they indicate deposition with little or no subsequent reworking by currents. The interaction of rate of subsidence and intensity of current action was such that for the most part the site of deposition was below the level of strongly agitated water.

The only fossils found widely in the Springer are Foraminifera which presumably were surface dwellers. Absent are the common bottom-dwelling forms such as brachiopods, mollusks, bryozoans, etc., which usually dominate the fauna in normal shallow marine waters. The absence of such a benthonic fauna is further evidence of the general stagnant unaerated environment of deposition.

The lack of calcite in the shales and the common occurrence of siderite indicates a low oxidation-reduction potential and a pH somewhat lower than normal (Garrels and Krumbein, 1952) such as is to be expected in a stagnant zone (Weeks, 1952). It has also been suggested by Weeks (1952) that the occurrence of primary chert as in some Springer shales may be due to the same environmental conditions.

The above generalizations concerning the environment of Springer deposition have their exceptions, the most important of which are found in association with the Springer oil reservoirs of the western part of the basin. These reservoirs constitute a distinctive facies in which nearly all of the features mentioned are absent or are reversed.

The sandstones are massive and perfectly winnowed of their clay matrix. A benthonic fauna is present and locally makes biostromal accumulations. Siderite is absent, but calcite occurs both as a cement and as oolites. Glauconite is a common authigenic mineral, and quartz overgrowths are common.

All of the features indicate deposition of the reservoir rocks in an agitated aerated oxidizing environment. It is apparent that these sediments accumulated in normal marine waters and well within the zone of wave and current agitation which operated on the detritus and modified it for a prolonged period of time. Deposition and subsidence was not so rapid but that the shallow water environment could leave its imprint on the sediment.

These areas in which the reservoirs are now found were likely topographic highs during the period of deposition; and on these highs the finer detritus was removed from the sands by wave or current action, and an aerobic environment maintained by the shallow agitated waters. As the oil field areas are now on or adjacent to large anticlines with structural relief of several hundred to several thousand feet, it is likely that the submarine topographic highs were maintained by early growth of the anticlines. That deformation took place during deposition is suggested by the common occurrence of slumped beds on the flanks of the anticlines.

The generally stagnant environment disappeared gradually and not at the same time in all places, but the post-Morrow formations show much less evidence of the stagnant marine environment of Springer deposition. Rather they were deposited in unrestricted shallow marine waters. A benthonic fauna consisting mainly of brachiopods, fusulinids, corals, Bryozoa, and gastropods occurs throughout the Dornick Hills, Deese, and Hoxbar formations. The generally light color of the sediments and the occurrence of red beds in the Deese formation indicate an oxidizing environment of deposition. The red beds, which are primary depositional red beds, are also evidence of a climate sufficiently warm and moist in the source area to produce red soils there.

Shallow agitated waters are indicated by the low amount of clay matrix in the sandstones and by the common occurrence of cross-bedding and ripple marks.

The overall rapid rate of accumulation of the Dornick Hills, Deese, and Hoxbar formations and the large number of fanglomerates which are present on the margins of the basin suggest that these formations are an example of most rapid, so-called "orogenic" sedimentation. The great thickness of sediment represents a corresponding great subsidence of the basin floor, and the fanglomerates are unmistakable evidence of sharp uplift on the basin's margins. The sandstones do not have the obvious evidences of very rapid deposition such as poor sorting and a large content of unstable minerals. However, poor sorting is indicated in a subtle way by the lack of hydraulic adjustment of the metamorphic accessory minerals; and the unstable "minerals" are present as detrital shale and limestone fragments. The abundant presence of shale and limestone grains can best be explained as the product of rapid erosion, short transport, and rapid rate of accumulation. Considering the well-sorted sedimentary detritus available in the source area, these sandstones are as "immature" both texturally and mineralogically as any arkose.

Insofar as sedimentary processes are concerned, these Ardmore Basin sediments differ little from the arkoses which directly overlie the Hoxbar formation. In each the essential feature is rapid deposition from intermittently active uplifts along the basin margins. The critical difference between the orthoquartzites of the Hoxbar formation and the arkoses of the Pontotoc formation is provenance, not type of sedimentation. The change in sediment represents no change in the nature of sedimentation, but only progressively deeper erosion on the borders of the basin cutting from the Simpson sandstones to the Precambrian granites. The same sedimentary processes which formed the Dornick Hills, Deese, and Hoxbar sediments would, if they had occurred in Cambrian time, have resulted in the deposition of a true tectonic arkose.

Source of the Conglomerates

In the conglomerates exposed on the south margin of the basin (the Jolliff and Bostwick members) the coarse detritus clearly originated from the Criner Hills. The conglomerates thicken and become coarser toward the Hills, and contain limestone pebbles only in the exposures near the Criner Hills. All of the formations now exposed in the Criner Hills, down to and including the upper part of the Arbuckle limestone, are represented in these conglomerates.

The chert which constitutes the largest part of the coarse material came from several formations. The Woodford chert was the chief contributor, but cherty zones occur in the Hunton, Viola, and Arbuckle limestones and at least locally supplied more than half of the chert pebbles.

The conglomerates of the western part of the basin are known from only a few scattered wells. However, they occur adjacent to the uplifted margins of the basin; and the older Paleozoic formations of these uplifts can be recognized in the pebbles which have been obtained in a few cores.

The Devils Kitchen conglomerate and other Deese conglomerates of the southeastern part of the basin contain no limestone pebbles by which the source formations may be recognized; nor, because of the unconformable Cretaceous cover, can they be seen in close proximity to their source area. However, these conglomerates are not different in kind than those of the southwestern flank of the basin. They are wedge-shaped basin margin deposits, and they probably originated from faulted uplifts either east or south of the present area of exposure.

This interpretation of the Devils Kitchen and related conglomerates differs from that of Tomlinson (1929, 1953). Because the conglomerates thicken and become coarser toward the southeast and because no limestone pebbles are present, Tomlinson maintained the source area must be to the southeast and could not have been the Arbuckle facies with its thick carbonate sequences.

However, as already pointed out (page 35), the apparent coarsening and thickening of the Devils Kitchen conglomerate toward the southeast is only a two dimensional component of the true relationship. There is no reason to assume the facies "strike" (Krumbein, 1952) to be perpendicular to the line of outcrop; it is in fact more commonly parallel, as for the Bostwick conglomerate of the opposite flank. The outward portions of alluvial fans may assume at times the shape of radiating and twisting fingers going out from the main mass. In such occurrences two dimensional outcrops may give any number of directional pictures. The direction of maximum increase in thickness and coarseness could as well be to the east, or possibly slightly north of east or directly south rather than to the southeast.

The absence of limestone pebbles may be due, as considered by Tomlinson, to the absence of limestone in the source area; but an equally satisfactory explanation is that the limestone pebbles were destroyed during sedimentation. The part of the Devils Kitchen conglomerate nearest the basin margin is not exposed; and as can be seen in the Bostwick conglomerate, limestone pebbles are confined to a narrow zone along the basin margin. Nearly all of the chert pebbles in the Devils Kitchen show a great deal of pre-depositional weathering; it would be highly unexpected to find any limestone pebbles capable of surviving the degree of weathering shown by the chert. If limestone pebbles are present, they would be expected only farther east in the upper part of the conglomerate wedge.

Without more detailed study it is not possible to trace the Devils Kitchen chert pebbles to a definite source. However, none of the distinctive chert lithologies of the Ouachita Mountains (Honess, 1923; Goldstein and Hendricks, 1953) were recognized in the Devils Kitchen pebbles.

Source of the Sandstones

In his basic paper on Ardmore Basin geology Tomlinson (1929) regarded the main sources of the sediments to be in highlands which were located some distance to the southeast of the basin, and nearly all writers since have followed this interpretation (van der Gracht, 1930; Kay, 1951).

The role of such a land area in the general sedimentation of southern Oklahoma is obvious (Miser, 1921) as all Paleozoic detrital units, with the exception of the Ordovician orthoquartzites and late Pennsylvanian and Permian arkoses, thicken and become coarser toward the southeast, and either disappear or grade into limestone toward the northwest.

However, the petrographic evidence from the Ardmore sandstones suggests that no single area could have supplied all the detritus for these rocks. The provenance of the sediments must necessarily have been complex as three distinctive petrographic facies have been recognized in the sandstones; these are: (1) the orthoquartzite assemblage of angular quartz and ultra-stable heavy minerals of the Springer sandstones; (2) the angular quartz, rock fragments, and metamorphic heavy minerals of the Dornick Hills, Deese, and part of the Hoxbar formation; and (3) the rounded quartz and heavy minerals, the detrital limestone, and the metamorphic heavy minerals of the outcropping Hoxbar sandstones.

The Springer was deposited in a relatively narrow trough across southern Oklahoma, and it is paleogeographically possible that Springer sediments may have come from several source areas. However, the homogeneity of the sandstones suggests that only one petrologic provenance was important.

The immediate source of the Springer sandstones was an area of sediments, but the indicated ultimate source seems to have been a metamorphic terrane. Although the sandstones are essentially concentrations of stable minerals, occasional metamorphic rock fragments and acid plagioclase grains in the main constituents, and traces of garnet and staurolite in the accessories suggest an area of predominantly metamorphic rocks for the original detritus.

With present information no definite conclusion as to the location of this area can be reached. The simple mineralogy, high degree of size-sorting, and apparent depth of water control of sedimentation suggest a source on the craton, or north side of the basin. However, no petrographically adequate provenance seems to be present in the area.

If the Springer is of Pennsylvanian age the Chester rocks are perhaps a possibility. They were, presumably, deposited over a larger area than that in which they are now preserved; but where these rocks are now known on the south flank of the Ozark uplift, and the north flank of the Anadarko basin they do not have the quantity of detrital sediments, particularly sandstone, which would be expected in the principal source of the Springer. And, of course, if the Springer is the lateral equivalent of these Chester sediments this possibility ceases to exist.

The other indicated possibility for a source area is to the southeast. The Stanley sediments are a quantitatively adequate source, and the graywackes of the Stanley could yield quartzose sands similar to those of the Springer after a moderate degree of weathering and size-sorting.

The Stanley rocks of the Ouachita Mountains were deposited over a much greater area than that over which they are now exposed. They represent sedimentation from what was probably the major paroxysm in the deformation of the Ouachita Mountains; and it could well have been that as the area of deformation expanded toward the interior of the continent, previously deposited Stanley rocks were uplifted to form the source of the Springer sediments. Possibly more than one cycle of erosion and deposition was involved.

The Springer unit as a whole thickens greatly toward the southeast, which is suggestive of a source in that direction, but the amount of sand does not show a similar increase. In fact, from the small area for which information is available the amount of sand appears to be less in the southeastern part of the basin than in the northwestern part. However, in the area of close subsurface control it is apparent that the distribution of sand was controlled more by bottom topography than by distance from source. Each sandstone member represents a regional tendency toward sand development, but within each unit variation is considerable, and the amount and kind of sand was apparently controlled by local environment.

In summary, the source of the Springer sediments is obscure and equivocal. The most likely provenance appears to be the Stanley sediments to the southeast of the Ardmore Basin, but the evidence is not substantial.

In contrast to the Springer, the later sediments show evidence of being a mixture of detritus from two or more source areas. The two provenances which can be most clearly recognized are the sediments of the local uplifts bordering the basin, and a metamorphic terrane which was probably on the general Red River uplift.

Several lines of evidence indicate that the sediments are of mixed origin. Among these are:

(1) The metamorphic accessory minerals are not in hydraulic adjustment with the rocks in which they occur. These minerals are poorly size-sorted, although the quartz and ultra-stable heavy minerals are excellently size-sorted. Also, the abundance and relative proportions of the metamorphic minerals vary greatly from place to place, suggesting that availability is the controlling factor in their occurrence.

(2) The same suite of metamorphic minerals occurs with two assemblages of quartz, tourmaline, and zircon—the angular assemblage of the Dornick Hills and Deese groups, and the well-rounded one of the Hoxbar group. All cannot have had the same source; the occurrence of the metamorphic minerals must be independent of one or both of the stable assemblages.

The contribution of the local uplifts is most obviously shown in the conglomerates; but the conglomerates do not bulk large in total volume, and the greatest contribution of these uplifts was to the finer detrital rocks. In general, the sediments portray the progressive unroofing of the marginal uplifts. The climax of this localized sedimentation came with the exposure of granite in the Arbuckle Mountains during Pontotoc sedimentation, but Springer and earlier Paleozoic sediments had been stripped off the uplifts earlier and incorporated into the Dornick Hills, Deese, and Hoxbar sediments.

The stratigraphic units which supplied the Dornick Hills, Deese, Hoxbar sandstones were chiefly the Simpson sands of mid-Ordovician age and the Springer sands of earliest Pennsylvanian age. The Simpson sands were an important source only for the Hoxbar formation, and then in only the area directly around the Arbuckle Mountains. Minor contributions from the Simpson are found in Dornick Hills sandstones around the Criner Hills, but in general the uplifts other than the Arbuckles do not have much Simpson sandstone or were not eroded to sufficient depth to expose the Simpson.

The presence of reworked Simpson sandstone is easily recognized by the extreme rounding of the quartz and heavy mineral grains. The similarity of the Hoxbar and Simpson sandstones can be seen in photomicrographs of thin sections, loose grains, or heavy minerals as in Figures 6 and 22, 9 and 24, and 10 and 35. Furthermore, grain size and sorting and the varieties of quartz and tourmaline match closely in the two formations. In addition to the similarity of the quartz and accessory minerals, the Hoxbar sandstones contain detrital limestone and chert grains which must have originated from a sequence of limestones such as those with which the Simpson sands occur. Some of the limestone grains can be matched with well known types of the Arbuckle Mountains; particularly distinctive are grains of the silty limestone of the Sycamore formation.

For the sandstones other than those of the Hoxbar around the Arbuckle Mountains the sands of the Springer formation seem to have been a major source. The stable constituents of the Dornick Hills and Deese sandstones are extremely similar to those of the Springer sandstones. The same varieties of quartz, tourmaline, zircon, and rutile seem to be present in all of these sandstones; and the grain shape and size distributions are very much alike (Figures 36 and 37; Tables 8 and 9).

Actually the similarities between the Springer sandstones and the Dornick Hills-Deese sandstones can be explained by assuming a continuation of sedimentation from the same source as for the Springer, as well as by reworking of previously deposited Springer sandstone—or by a combination of the two. However, several things suggest that most of the sediment is largely reworked Springer directly from the flanks of the basin. One is the large amount of shale fragments that is found in the Dornick Hills-Deese sandstones; this is not likely to have been transported more than very short distances. Several paleogeographic facts also support this interpretation; among these are: (1) the Arbuckle Mountains were eroded into the Ordovician strata by the beginning of Hoxbar sedimentation; and hence, all younger Paleozoic sediments must have been stripped off before Hoxbar time and have contributed to the sediments of the Ardmore Basin; (2) drilling has shown that a number of large anticlinal structures in the southern and western parts of the basin, including the structures of Velma, Fox-Graham, Healdton, and Hewitt oil fields, were eroded of part or all of their lower Pennsylvanian sediments before they were buried by upper Deese, Hoxbar, or Pontotoc sediments; the detritus from these uplifts must necessarily have contributed to the Ardmore sediments; and (3) the occurrence of conglomerates at a number of levels in the Deese and Dornick Hills groups show that the basin margins were land areas supplying sediment during Dornick Hills-Deese sedimentation.

The volume of the Dornick Hills and Deese sediments in the Ardmore Basin is not more than 350 cubic miles. This can readily be accounted for by erosion of the highlands immediately surrounding the basin, as a conservative estimate of the land areas supplying sediment indicates a minimum of 500 square miles.

The contribution of the metamorphic terrane is most clearly shown by the abundant accessory minerals of metamorphic origin in many of the samples, but is also indicated by small amounts of metamorphic rock fragments.

For the exposed sediments the metamorphic rocks were not a quantitatively important source, and many samples show little or no evidence of any source other than the earlier sediments from the local basin margin uplifts. In the samples from wells in the western part of the basin the metamorphic source is more obvious, and in this part of the basin apparently played a bigger part in the overall sedimentation than it did around Ardmore. The greater influence of the metamorphic terrane in the western part of the basin is perhaps most clearly shown in the distribution of the heavy minerals (Table 5). Metamorphic minerals, particularly garnet, are more abundant in the western samples, are more uniformly distributed, and occur in greater variety.

This distribution of accessory minerals indicates that their source was probably to the west or southwest of the Ardmore Basin. Such a source is also suggested by the occurrence of chloritoid, metamorphic type chlorite, and clinozoisite at lower stratigraphic levels in the western part of the basin.

The specific area from which these minerals came was most probably the Red River uplift. This area was emergent during much of Pennsylvanian time, and Precambrian medium rank metamorphic rocks are now overlain directly by upper Pennsylvanian sediments (Flawn, 1956). These basement rocks are the only petrographically adequate source now known for the metamorphic contribution to the Pennsylvanian sediments of the Ardmore Basin. It is possible that low and moderate rank metamorphic rocks may occur in the buried portions of the Wichita Mountain trend and have been a source. The exposed Wichita Mountains are largely granite with some gabbro, but a small area of high rank metamorphic rock (Meers quartzite) is known, and drilling has disclosed the presence of igneous flows and low rank metamorphic rocks on the Wichita trend just a few miles to the west of the area studied here.

In addition to the two relatively near-by recognized sources other areas may well have contributed minor amounts of detritus to the basin. The Wichita Mountains were uplifted in early Pennsylvanian time, and contributed arkose to Deese equivalents in the western part of the Anadarko Basin. The presence of occasional flakes of fresh biotite in some samples from the western part of the basin suggests a source in the Wichita Mountain rocks, but evidence of such a source is scant east of the Velma uplift, and it was probably not quantitatively important.

The general land area which supplied the bulk of eastern Oklahoma sediments was probably cut off from the Ardmore Basin by the uplift of the Arbuckle Mountains, but for general Pennsylvanian sedimentation there were source areas to the southeast of the present Ouachita Mountains, in the Ozark Mountains, in the Trans-Continental arch of the North Central United States, and in the southern Rocky Mountains, all of which could have been the origin of detritus not readily distinguished in small amounts.

A particular attempt was made to recognize a contribution of detritus from the Ouachita Mountains, but this could not be done. No positive evidence that the Ouachitas were or were not a source area for Pennsylvanian sediments was found, but there are several lines of negative evidence that indicate the Ouachita Mountains were probably not a major source area. Among these are:

(1) No grains or pebbles of Ouachita rocks could be recognized in the Pennsylvanian sediments of either the Ardmore Basin or the McAlester Basin north of the Ouachitas.

(2) The Jackfork sandstone, which is the main body of sand in the Ouachita Mountains and would necessarily be a major source of detritus, approaches being an orthoquartzite and has the characteristic orthoquartzite mineral assemblage of ultra-stable minerals. It has a simpler mineralogy and a lower degree of size sorting than any of the Pennsylvanian sandstones for which it could conceivably have been a source. No combination of Ouachita rocks could apparently have furnished the mineral composition of the Ardmore Basin sediments.

(3) No unconformities or orogenic conglomerates which can be attributed to uplift in the Ouachita Mountains are found in the Pennsylvanian strata bordering the mountains.

(4) The Ouachita Mountains were themselves receiving sediments at least until near the beginning of Deese deposition. The Atoka formation is the same age as much of the Dornick Hills group, and hence, can hardly have been a source of the Dornick Hills. As the Deese sandstones are petrographically indistinguishable from those of the Dornick Hills (both differ from the Atoka) one may doubt that the Ouachita Mountains were the source of either.

(5) If the Ouachita Mountains were a major source area similar sediments would be expected in the McAlester Basin to the north, and the Ardmore Basin to the west of the mountains. This is not, however, what is found. The McAlester Basin sandstones are generally dissimilar to those of the Ardmore Basin. In particular, they are characterized by abundant mica which is present in only traces in the Ardmore Basin sandstones.

These observations are perhaps not sufficient to rule out the possibility of a Pennsylvanian Ouachita uplift, but recognizable sources of Pennsylvanian sediments seem to be adequate, and in the absence of positive evidence one has little reason for assuming the Ouachita Mountains were a source area.

RESERVOIR PETROLOGY OF THE SPRINGER SANDSTONES

Introduction

It has been pointed out throughout the description of the sediments that the Springer reservoir sandstones of the Ardmore basin are petrographically distinctive. They illustrate with a high degree of perfection the "specific reservoir" concept as developed by Krynine (1948; 1951a; 1951b) and the differentiation of a reservoir and source facies within a sedimentary basin as described by Weeks (1952).

General Character of the Springer Sediments

The Springer formation consists of about 4,000 feet of sediment, most of which is dark gray shale. Several sandy zones occur in the upper part of the unit, and in these are found the oil reservoirs. Laminae or bands of siderite occur throughout the shale, and a trace of limestone is present in association with the oil reservoirs.

The shale is characteristically very fine-grained and lacking in detrital constituents other than clay minerals. Illite is the dominant clay mineral, but kaolinite is present in all samples and montmorillonite in most. Since the composition of the sandstones suggests that the Springer is an accumulation of stable minerals, the montmorillonite probably represents volcanic ash falls; there is evidence of volcanic activity in the Ouachita Mountains in late Mississippian time. Siderite is by far the most abundant authigenic mineral, but pyrite is common, and much of the shale is slightly siliceous. The dark color of the shale is due to abundant carbonaceous material.

Sandstone makes up less than fifteen per cent of the Springer formation, but locally in the area of Springer oil production the upper half of the unit is approximately fifty per cent sandstone. The sandstones occur in four or five zones which have at least local continuity. These sandy zones consist of a large number of incompletely coalescing sand bodies which change thickness and character over a distance of a mile or less.

The sandstones range from thin-bedded to massive, and the massive sandstones may be either clean quartz sandstones or clayey sandstones with a uniformly distributed matrix. In grain size they are either fine- or very fine-grained, and the sand fraction is uniformly well-sorted. Quartz and clay minerals are the only abundant detrital constituents; and the ultra-stable tourmaline, zircon, and rutile comprise nearly the entire heavy mineral assemblage. Bonding of the rocks is either by authigenesis of the clay minerals or by chemical cements. Quartz, calcite, and siderite are found as cements; however, quartz and calcite are found only in the more quartzose sandstones. In an occasional sand body, or part of one, bonding is virtually absent. Other than the cements the only authigenic mineral is glauconite; it is a minor constituent of many of the sandstones.

Limestone is a minor part of the Springer formation, and has only been found in a few thin beds in the oil field area of the Ardmore basin. Two types of limestone are found: one is an oolitic limestone in which the oolites are dark gray and have nuclei of quartz grains that are the same size as the grains in the associated sandstones; the second type is a bioclastic limestone made up of fragments of brachiopods, bryozoans, crinoids, and algae.

The fauna of the Springer formation is sparse, and except in a few small areas is completely lacking in the normal bottom forms of a shallow marine sea. A long range planktonic microfauna is present, and collections of goniatite casts and sharks' teeth have been made. In addition to the fauna occasional fragments of plant fossils are found.

Petrography of the Springer Reservoirs

The sandstones which make up the Springer reservoirs are a distinctive lithologic type that can be readily identified by petrographic criteria. In most of their characteristics they are not a completely separate population, but rather are one extreme of the general range of distribution of the individual characteristics. In all characteristics the Springer sandstones have a considerable range of values, but reservoirs are found within only a small part of that range. The better the reservoirs the more distinctive they are.

The more obvious characteristic features of the reservoirs are as follows:

(1) The sandstones are highly quartzose. Ordinarily quartz constitutes 95 per cent or more of the detrital constituents, but in the better reservoirs the percentage may approach 100. In particular, little or no clay matrix is present in the better reservoirs.

(2) As a group the oil producing sandstones are coarser than the non-producing ones. Nearly all of the better reservoir sandstones are fine-grained whereas the bulk of the Springer sandstones are very fine-grained (figure 36). No significant difference in size sorting was found.

(3) Bedding of the reservoirs tends to be massive, and the sandstone occurs in unbroken sequences as much as a hundred feet thick. The reservoirs grade laterally into thin-bedded sandstones which commonly show depositional slumping.

(4) The reservoir sandstones contain a higher proportion of heavy minerals than do the other sandstones. In a few samples selected size fractions had as much as five per cent heavy minerals, whereas the non-reservoir sands characteristically had less than one-half per cent. The heavy mineral assemblage of the reservoirs is also more restricted; semi-stable minerals such as garnet and staurolite occur in small amounts in the ordinary Springer sandstones but are absent in the reservoir sandstones.

(5) With the exception of the outcropping Target limestone (Bennison, 1954) the only limestones known in the Springer occur in association with the reservoir sandstones. The limestone is either bioclastic or oolitic. On the crest of the Sholom Alechem anticline the Sims sandstone member consists in large part of fossil fragments, and in most of the coarser-grained sandstones limestone fragments are common.

(6) Siderite which is a common cementing mineral in the Springer as a whole is absent in the reservoir facies. On the other hand, calcite which is a rare mineral in the normal Springer lithology, is common, both as clastic grains and cement, in the reservoir sandstones.

Distribution of the Reservoir Sandstones

All of the Springer oil fields occur on top or on the flank of moderately large to large anticlinal structures. The chief producing area is along the uplift upon which the Fox-Graham and Velma fields occur; other important fields are on the large Sholom Alechem Anticline, and on the west plunge of the Arbuckle Anticline and the Pauls Valley Uplift. Although the sands are lenticular, there has been no important production from isolated sand bodies independent of a large uplift.

On these anticlines the character of the reservoir seems to be related with structural position, with the best reservoirs found in the highest positions structurally. The reservoirs change down dip by the intercalation of thin shale partings which become progressively more abundant. The downdip limit of the reservoir is poorly defined but is marked by gradual reduction in the permeability of sands. There is no edge water; and, hence, no water drive in any of the oil accumulations. The relationship between structural position and quality of the reservoir has been noted by Davis (1951), and the data in Table 17 are taken from his work.

TABLE 17
RELATIONSHIP OF STRUCTURAL POSITION AND RESERVOIR PROPERTIES IN THE SPRINGER RESERVOIR OF VELMA

Well	Subsea Top 1st Sand	Gross Ft. of Section	Net Ft. Oil Pay	Avg. Perm. Md.	Avg. Poros. %
A	-3722	335	238	300	21
B	-4022	315	194	190	19
C	-4871	334	212	70	16
D	-5398	292	172	40	14
E	-5892	302	65	31	12

In five wells which cover a range of slightly more than 2,000 feet of structural relief the average permeability decreases from 300 to 31 millidarcies, the porosity from 21 to 12 per cent, and the number of feet of productive sand from 238 to 65 feet. Similarly detailed data are not available for the other Springer fields, but the evidence all supports the interpretation of a general downdip deterioration of reservoir quality. In all of the fields studied, the coarsest and most quartzose samples were from the structurally high wells, and there was progressive decrease in grain size, and increase in clay content in structurally lower wells. Electric log cross-sections perpendicular to the strike characteristically show a breaking up of the massive up-dip sand bodies, and a decrease in the amount of net sand and in porosity as the units are followed down dip.

At Velma (data for other fields not available) the oil systematically becomes heavier and more viscous down dip, and the edge of the field is in part controlled by the occurrence of oil too heavy and viscous to produce economically. The systematic change in the character of the oil is in keeping with the change in rock character, and can possibly be attributed to the change in clay content of the reservoir (Chilingar, 1955), or to differences in the physical-chemical environment of origin.

Origin of the Reservoir Sandstones

The contrast between the reservoir sandstones and the remainder of the Springer is largely the contrast between deposition in relatively highly agitated waters and deposition in relatively quiet waters. In the reservoirs the winnowed character of the sands, the massive bedding, the coarser grain size, and the placer-like concentration of heavy minerals attest the action of strong currents whereas the presence of a benthonic fauna and of oolites indicates aerated agitated waters of normal salinity and pH.

On the other hand, the bulk of the Springer is an almost perfect example of what Weeks (1952) has termed a "source facies." The unwinnowed or thin-bedded sandstones suggest deposition in relatively quiet waters while the absence of a benthonic fauna and the presence of a fauna of floaters and swimmers indicates that the bottom environment was probably stagnant. The absence of calcite in these sediments, the abundance of siderite, the occasional occurrence of pyrite and chert, and the abundance of carbonaceous material form associations characteristic of a narrow range of physico-chemical conditions. Apparently the only important variable influencing the deposition of calcite in a marine environment, either physico-chemically or biochemically, is the pH of the environment; and the failure of calcite to be deposited in an environment in which chemical sedimentation was taking place probably reflects a pH less than that of normal open-water seas (7.5-8.5) and probably close to 7.0 At a pH of 7.0 siderite will be deposited in an environment with an oxidation potential from 0.0 to -0.2 and pyrite below -0.2 (Krumbein and Garrels, 1952). Thus, the indicated bottom environment is one which is only slightly alkaline, or possibly even neutral, and which is moderately reducing most of the time but occasionally is strongly reducing. As the Springer fauna indicates that the surface waters were oxygenated and of normal salinity, it is probable that Springer deposition was in a basin in which circulation was partially restricted by a sill.

As Weeks (1952) has pointed out, this can be an ideal condition for the generation of petroleum.

Deposition of the reservoir facies within such a basin requires that there be local variation in the relationship between the surface of deposition and the depth to which current and wave agitation extend. Where deposition was at a depth of slight or no agitation, near-stagnant or stagnant conditions existed; but at the same time deposition on isolated topographic highs could be on strongly agitated bottoms with an oxidizing environment. On these highs the physico-chemical environment would be that of the open circulation marine environment, that is, mildly oxidizing and alkaline and with normal marine salinity. Under these conditions calcite could be deposited either biochemically or physicochemically; and at periods when sand was entering the basin, these highs would be sites where the currents would winnow the finer grains from the detritus and leave on the highs the relatively more coarse sand which accumulated as the massive quartzose reservoir rocks.

Such topographic highs may be the random irregularities of any sea bottom; and, thus, the reservoir environment may have existed briefly at almost any place within the basin.

However, the critical factor seems to be that structural deformation was going on during sedimentation. The fact that the major Springer oil reservoirs occur in association with large structural highs indicates that the variation of the surface of deposition was the topographic reflection of structural features. Further evidence that the structural features controlled the surface of deposition and that the anticlines were growing during sedimentation is the common occurrence of slumping in the sediments along the flanks of the anticlines (p. 52). At least along the flanks of the Fox-Graham-Velma uplift the slumping has consisted of relatively large displacements and strongly suggests the existence of slopes along which the unconsolidated sediments could slide.

During deposition these embryonic anticlines or other uplifts were probably areas of lesser subsidence, and were sites in which any detritus would be subjected to more intense and especially more prolonged current agitation than the surrounding areas, and, hence, would accumulate sediment which could preferentially store oil (Krynine, 1951).

Conclusions

Springer sedimentation was for the most part in a basin of restricted circulation where the physical-chemical environment was near the optimum for the generation of petroleum. The Springer sandstone reservoirs were deposited on local sites of a quite different environment—one of agitated and oxygenated waters.

The localizing factor for the major reservoirs seems to have been in all instances a structure which was growing during sedimentation and was, hence, able to maintain shallow water or to hold a shoreline for a considerable period of time. The critical factor in the formation of the Springer reservoir sandstones was not the environment in which they were deposited, but rather the length of time they were exposed to that environment.

These uplifts are not structural traps in the usual sense, but have localized the accumulation of oil through being the sites of the deposition of potential reservoir sandstones. Hence, not all anticlines are equally probable oil pools; only those anticlines with an early tectonic history are likely to have adequate reservoirs; anticlines formed entirely during the Arbuckle (late Pennsylvanian) orogeny cannot be expected to have major Springer oil accumulations.

Similarily, any other type of "trap" which is independent of a local uplift that was active early is not a favorable area for oil accumulation. Truncation of a sand body is not in itself likely to form a trap. Only where the truncation cuts a zone of favorable facies is there likely to be an accumulation, and in such instances the truncation is strictly a factor of secondary importance. The main effect of the truncation is to decrease the size of the available trap.

While certain conventional types of oil traps are not likely to be found in the Springer sandstones, it is not necessarily true that the only Springer oil fields will be found on anticlines. As the early tectonic uplift, not the anticline, plays the important role in the formation of a reservoir, any faulted monocline, horst, or unclosed anticline is a potential site of a reservoir if it were being formed during Springer sedimentation.

Since the presence of a reservoir is related to the presence of a growing structure, the size and position of the reservoir are likely to be related to the magnitude of the uplift. The larger the uplift and the greater the amount of structural relief, the more likely are the sandstones to have been exposed to prolonged current and wave agitation. Also, where favorable sandstones occur far down on the flanks of a large uplift, the uplift was presumably a large topographic feature during deposition of the sand. Where the reservoir sands are limited to a flank occurrence, such as at Velma, it is probable that the extent of the sand is not controlled by truncation alone but that the sands were not deposited completely across the topographic feature which existed during sedimentation.

Presumably the oil in the Springer is indigenous. The Springer shales which surround the reservoir sandstones seem to be an ideal source facies, and there is no apparent need to look further for the source of the Springer oil. Permeability is a locally developed feature of the Springer sandstone zones, and anything but local migration of the oil does not seem to have been possible.

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APPENDIX

LOCATION AND STRATIGRAPHIC POSITION OF SAMPLES

Sample	Stratigraphic Position	Location
DH-B-13	Bostwick cg. Dornick Hills gr.	NW NE NW 27-4S-1E
D-DK-25	Devils Kitchen cg. Deese fm.	NE NE SW 4-6S-2E
D-DK-29	do.	SW SW SW 13-6S-2E
D-32	Middle Deese	SW NW NW 33-5S-2E
H-34	Basal Hoxbar	NE SW NE 17-5S-2E
DH-36	Big Branch fm. Upper Dornick Hills (?)	SE SW SE 27-4S-1E
D-DK-40	Devils Kitchen cg. Deese fm.	SW SW NW 18-4S-2E
DH-B-43	Bostwick cg. Dornick Hills gr.	SW SW SW 7-4S-2E
S-RC-45	Rod Club ss. Springer fm.	SE SE NE 12-4S-1E
S-RC-46	do.	do.
S-LA-48	Lake Ardmore ss. Springer fm.	NE NE SE 1-4S-1E
S-P-49	Primrose ss. Springer fm.	do.
D-70	Lower Deese fm.	Schermerhorn #3, Calvery, 3743-3747′, NW SE SW 2-2S-3W
D-72	Fusulinid sand Deese fm.	Schermerhorn #9 Sparks, 2115′, C SL SW SW 30-2S-2W
D-75	Lower Deese fm.	Schermerhorn #1 Morris, 3201′, SW SW NW 2-2S-3W
S-77	Sims ss. Springer fm.	Gulf #3 Pruitt SW SE SE 29-2S-3W
S-78	do.	Gulf #9, Lindesmith SE NW NW 33-2S-3W
S-80	Humphreys ss. Springer fm.	Seaboard #2 Carpenter, 6579′, SE SW SW 11-2S-3W
H-82	Lower Hoxbar fm.	NW NW NE 7-5S-2E
H-84	Lower Hoxbar fm.	SE NW SE 8-5S-2E
H-Z-90	Zuckermann ss. Hoxbar fm.	NE NE NW 16-5S-2E
D-N-98	Natsy ss. Deese fm.	SE NW SE 17-5S-2E
S-S-104	Sims ss. Springer fm.	Continental #1A Martin SW SW NW 6-2S-3W
D-112	Upper Deese fm.	SW SW SE 20-5S-2E
D-122	Middle Deese fm.	NW NE NW 4-6S-2E
D-130	Lower Deese fm.	NW NW NE 9-6S-2E
DH-140	Big Branch fm. Upper Dornick Hills	NW NE NE 28-4S-1E
DH-B-146	Bostwick cg. Dornick Hills gr.	SE NW NE 28-4S-1E
D-151	Middle Deese fm.	C SL 16-4S-1E
S-165	Lake Ardmore (?) ss. Springer fm.	NW SE SE 33-2S-1E
S-LA-174	Lake Ardmore ss. Springer fm.	NW NW NE 11-3S-1W
S-RC-175	Rod Club ss. Springer fm.	SE NW SE 2-3S-1W

Sample	Stratigraphic Position	Location
D-176	Lower Deese fm.	NE SE NE 8-3S-1E
S-O-178	Overbrook ss. Springer fm.	NE SE SE 17-3S-1E
D-181	Upper Deese fm.	NE NW SE 14-4S-1E
DH-186	Big Branch fm. Upper Dornick Hills	NE NW SE 11-4S-1E
S-LA-190	Lake Ardmore ss. Springer fm.	SW NE SE 2-4S-1E
S-O-195	Overbrook ss. Springer fm.	SW NW NE 2-4S-1E
S-RC-196	Rod Club ss. Springer fm.	SW NE SW 35-3S-1E
G-198	Upper Goddard fm.	SW SW NE 35-3S-1E
D-207	Lower Deese fm.	NE SE SE 28-3S-1E
D-DK-214	Devils Kitchen cg. Deese fm.	SW SW SW 13-6S-2E
DH-B-216	Bostwick cg. Dornick Hills gr.	NE NE NW 25-6S-2E
D-218	Upper Deese fm.	SW SE NE 21-4S-2E
H-232	Lower Hoxbar fm.	SE SE SE 8-5S-2E
D-234	Middle Deese fm.	SW SE SE 10-3S-2E
H-236	Zuckermann ss. Hoxbar fm.	SE NW NE 16-5S-2E
H-239	Upper Hoxbar fm.	NW NW SW 15-5S-2E
DH-B-240	Bostwick cg. Dornick Hills fm.	SW SE SE 34-4S-1E
S-O-242	Overbrook ss. Springer fm.	NE NE NW 6-6S-2E
Si-245	Bromide ss. Simpson gr.	U. S. Highway 77, So. flank Arbuckle Mountains
Si-246	Tulip Creek ss. Simpson gr.	do.
Si-247	Tulip Creek ss. Simpson gr.	U. S. Highway No. 77, So. flank Arbuckle Mountains
Si-248	McLish ss. Simpson gr.	do.
Si-249	McLish ss. Simpson gr.	do.
Si-250	McLish ss. Simpson gr.	do.
Si-257	2nd Bromide ss. Simpson gr.	Skelly #1D Frensley B., 6684-6693′, C SW SW 25-1S-5W
Si-261	2nd Bromide ss. Simpson gr.	Skelly #4 Robberson, 7180-7194′, 35-1S-5W
D-272	Upper Deese fm.	Skelly #2 Hinkle, 2777-2793′, 32-1N-3W
D-278	Tussy ss. Deese fm.	Skelly #1 Johns, 3606-3610′, 19-1N-3W
D-284	Fusulinid ss. Deese fm.	Skelly #3 Gross, 2605-2617′, 32-1N-3W
D-285	Lower Deese fm.	Texas #2 Denny NE SE NE 1-2S-4W
D-286	Lower Deese fm.	Stanolind #1 Mills, 5332-5348′, 17-2S-3W
H-289	Lower Hoxbar fm.	Skelly #1 Humphries, 1815-1833′, SW SW SW 13-1S-5W
H-290	do.	Skelly #1 Humphries, 1975-1893′, SW SW SW 13-1S-5W

Sample	Stratigraphic Position	Location
D-292	Upper Deese fm.	Skelly #1 Humphries, 2628-2643', SW SW SW 13-1S-5W
D-305	Tussy ss. Deese fm.	Skelly #1 Baker, 3272-3290', SW NW NE 25-1S-5W
S-H-401	Humphrey ss. Springer fm.	Stanolind #1 Johnson, 5633-5664', 4-2S-3W
S-G-402	Goodwin ss. Springer fm.	Stanolind #1 Johnson, 6402-36', 4-2S-3W
S-403	Sims (?) ss. Springer fm.	Cumbic and Collins #1 Luster, 850-890', 17-1S-2W
S-H-417	Humphrey ss. Springer fm.	Magnolia #7 Ida Bumpas, 4871-4881', 33-1S-3W
S-S-423	Sims ss. Springer fm.	Magnolia #7 Ida Bumpas, 5265-5275', 33-1S-3W
S-S-424	Sims ss. Springer fm.	Magnolia #7 Ida Bumpas, 5276-5286', 33-1S-3W
S-G-429	Goodwin ss. Springer fm.	Skelly #E-4 Frensley, 3195-3212', NW NW SW 24-1S-5W
S-S-433	Sims ss. Springer fm.	Skelly #1 Romina, 3900-3914', SE SE SE 14-1S-5W
S-S-437	Sims ss. Springer fm.	Skelly #1 Baker, 4162-4182', SE NW SE 25-1S-5W
S-444	Sims (?) ss. Springer fm.	Skelly #32 Wirt Franklin, 3765-3777', NE SE SE 34-1S-5W
S-S-455	Sims ss. Springer fm.	Sun #8A Robison, 4973', SE NE SE 3-2S-4W
S-S-457	do.	Sun 8A Robison, 5012', SE NE SE 3-2S-4W
S-S-459	do.	Sun #8A Robison, 5049', SE NE SE 3-2S-4W
S-S-462	do.	Sun #8A Robison, 5122', SE NE SE 3-2S-4W
S-S-464	do.	Sun #8A Robison, 5139', SE NE SE 3-2S-4W
A-502	Lower Atoka fm.	3-3N-19E Latimer Co., Okla.
J-507	Jackfork ss.	6.8 mi. east of Antlers, Okla. State Hw. 3, Pushmataha Co., Okla
St-508	Upper Stanley fm. gp	31-4S-21E McCurtain Co., Oklahoma
St-510	Upper Stanley fm. gp.	31-4S-20E Pushmataha Co., Okla.
J-515	Lower Jackfork ss.	2-3N-19E Latimer Co., Okla.
St-517	Upper Stanley fm. gp.	1.5 mi. north of Moyers, Pushmataha Co., Okla.

A

B

C

D