Geology and Ore Deposits of the Picher Field Oklahoma and Kansas

GEOLOGICAL SURVEY PROFESSIONAL PAPER 588

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By EDWIN T. McKNIGHT *and* RICHARD P. FISCHER

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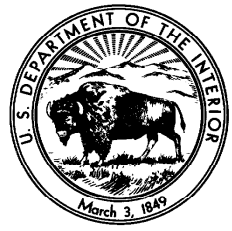
A discussion of one of the world's great mining fields—its geology, mining history, and potential



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GEOLOGY AND ORE DEPOSITS OF THE PICHER FIELD, OKLAHOMA AND KANSAS

By EDWIN T. MCKNIGHT and RICHARD P. FISCHER

ABSTRACT

The Picher field, one of the great mining districts of the world, has yielded more than a billion dollars' worth of zinc and lead since the first ore was marketed in 1904. The field, which straddles the Oklahoma-Kansas State line just a few miles west of the Missouri State line, was the last discovered and westernmost subdivision of the Tri-State mining region. The main part, about 8 by 9 miles, lies in flat prairie land west of Spring River at the eastern border of the Osage Plains. There are some outlying mines a few miles west or southwest of the main area.

The rocks exposed at the surface in the mining field encompass a relatively thin interval of Mississippian and Pennsylvanian strata that are nearly flat lying. Older rocks are cut in the mine workings or are exposed in the Wyandotte quadrangle south of the mining area, and still older strata have been intersected by deep drilling within the mining field.

Precambrian granite and possibly igneous flow rocks have been reached in 11 drill holes in and adjacent to the mining field at depths ranging between 291 and 1,942 feet. The Precambrian is overlain, with profound unconformity, by Upper Cambrian and Lower Ordovician strata to a maximum thickness of 1,325 feet, consisting dominantly of dolomite but with some sandstone and sparse shale. As the Precambrian terrain had been eroded to a landscape of sharp relief before burial by the Paleozoic strata, the earliest of these strata pinch out on the flanks of the buried granite ridges and peaks and are overlapped by higher formations. Formations that have distinguished in the well cuttings include, from the base up, the Lamotte Sandstone, 12–50 feet; Bonneterre(?) Dolomite, 45–75 feet; Davis Formation, 110–120 feet; Eminence Dolomite (possibly including thin Potosi Dolomite at base), 135–157 feet; Gunter Sandstone Member of Van Buren Formation (marking the base of the Ordovician), 20–40 feet; upper part of the Van Buren Formation and Gasconade Dolomite, 240–300 feet; Roubidoux Formation, 105–190 feet; Jefferson City Dolomite, 270–340 feet; and Cotter Dolomite, 143–183 feet. In two drille holes that hit a peak or high ridge in the buried Precambrian topography, the Cambrian and Ordovician formations are missing and Mississippian strata rest directly on the Precambrian.

The deepest surface exposures in the quadrangle south of the mining district have cut only 26 feet below the top of the Cotter Dolomite. The top part of this formation is a medium- to fine-grained gray to brownish-gray dolomite, slightly sandy in certain thin layers and containing a little chert of various colors, commonly oolitic.

Unconformably overlying the Cotter is the Chattanooga Shale of Late Devonian and Mississippian (Kinderhook) age. Most of it is black fissile shale, but locally near the top it is gray, nonfissile, and slightly sandy. At one locality on Buffalo Creek, the shale has at its base a white sandstone, 1–4 inches thick, which is probably equivalent to the Sylamore Sandstone Member of the Chattanooga in adjacent areas of Missouri and Arkansas. The best exposures of the Chattanooga were along the Neosho River in the southern part of the quadrangle and are now submerged, along with the underlying Cotter, beneath the Lake O' the Cherokees. Former exposures and well logs indicate a thickness of 20–67 feet in the southern part of the quadrangle, but the formation thins northward and disappears at about the latitude of the mining field.

The Chattanooga Shale is overlain disconformably by the Boone Formation, of Mississippian age. The Boone is composed of fossiliferous limestone, cotton rock, and chert, the latter occurring as nodules in the limestone and as interbeds which range to as much as 60 feet in thickness. The formation is 350–400 feet thick in the mining field where it is the host rock for most of the ore deposits. The Boone Formation is in this report subdivided into seven members: St. Joe Limestone Member, Reeds Spring Member, Grand Falls Chert Member, Joplin Member, Short Creek Oolite Member, Baxter Springs Member, and Moccasin Bend Member. In the southern part of the Wyandotte quadrangle the base of the Reeds Spring Member is distinguished by huge crinoidal bioherms several hundred feet across and as much as 100 feet thick. There is a disconformity at the base of the Short Creek Oolite Member. The lower part of the Boone is classed in the Osage Series, and the Short Creek Oolite Member constitutes the base of the overlying Meramec Series.

The Boone is overlain, with probable conformity, by the Quapaw Limestone. This formation is best known from a locality near the abandoned town of Lincolnville, where about 25 feet of the limestone is exposed. It is also present in the subsurface of at least a part of the mining field. It is a medium- to coarse-grained gray to brownish-gray crinoidal limestone containing a fauna of Warsaw age.

The Chester Series, composed of three conformable formations, is disconformable on the Boone Formation and presumably also on the Quapaw Limestone. Locally, its basal part filled sinkholes in the Boone. Its basal formation, the Hindsville Limestone, 0–85 feet thick, is composed of fine- to coarse-grained fossiliferous limestone containing minor oolite, lenticular sandstone, and chert. A basal conglomerate of chert pebbles in limestone is locally present. The Hindsville grades upward with indefinite intertonguing boundary into a sandier facies, the Batesville Sandstone, which, however, contains more limestone than sandstone, and has considerable oolite and green to gray shale. The Batesville thickness is 0–70 feet. Both Hindsville

and Batesville have been locally mineralized in the mining field. The Fayetteville Shale is the top unit of the Chester in the southern part of the Wyandotte quadrangle, but it has been removed in the mining field by erosion beneath the unconformity at the base of the Pennsylvanian, except where locally preserved in at least one or possibly more structural basins of small extent. It is a black to greenish fissile shale with some interbedded fine- to coarse-grained crinoidal limestone and some oolite, totaling 0–70 feet in thickness. The limier phases are very fossiliferous.

The Hale Formation overlies the Fayetteville unconformably and is preserved only in and adjacent to the Seneca synclinal graben in the southern part of the quadrangle. The most complete exposures, showing a maximum thickness of 83 feet, have been drowned by Lake O' the Cherokees. The formation comprises interbedded sandstone, shale, dark fossiliferous limestone, oolite, and one thin bed of "sandy iron ore." The sandstone is interbedded with shale in the lower half of the formerly exposed section and with limestone and oolite in the upper half. Both the sandstone and limestone are heavily pigmented with dark bituminous material, possibly oily, and the lower sandstones contain much carbonized wood and other fossil plant material.

The Krebs Group, of Pennsylvanian (Des Moines) age, is unconformable on all underlying formations. It is the surface rock throughout most of the mining field. Several formations have been recognized within the Krebs in areas south of the Wyandotte quadrangle, but within this quadrangle the dominant facies of all formations is black to gray fissile shale, which contains two named sandstone members and a little additional sandstone of irregular and discontinuous occurrence. The formations, as recognized and mapped by Branson, are, from the base up, the Hartshorne Formation, McAlester Formation containing Warner Sandstone Member at its base, the Savanna Formation, and the basal Bluejacket Sandstone Member of the Boggy Formation. A few clay-ironstone concretionary zones, thin limy beds, and thin coal seams occur at various levels, and there is a thin discontinuous chert pebble conglomerate at the base of the group. The Warner Sandstone Member, 0–20 feet thick, lies 0–50 feet above the base of the Krebs. The Bluejacket Sandstone Member, only the basal 60 feet of which remains, is about 200 feet above the base of the Krebs. It caps the highest hill in the mining field and also occurs in a synclinal graben trough that crosses the field. Except for showings of ore minerals, the Krebs Group is not mineralized but forms an impervious cap to the mineralized ground.

The unconformity at the base of the Krebs truncates underlying formations from south to north in the Wyandotte quadrangle, from the Fayetteville Shale at the south to the upper part of the Boone in parts of the mining field. The erosion cycle terminating with the Krebs deposition was marked not only by surficial erosion but also by underground solution of the susceptible limestones to form caves. These caves were filled during the initial stages of the Krebs inundation by shaly sandstone and black fissile shale carrying plant remains, and particularly seeds of *Cordiocarpus lineatus* Lesquereux, which have been carbonized and pyritized to preserve delicate details of their original form. Most of the filled caves revealed by mining operations in the mining field are in the Boone and particularly in the Joplin Member, perhaps 200 feet below the plane of unconformity and probably well below the ground-water table of that time. These caves follow the bedding and may be several hundred feet across in horizontal extent, but only a few feet high. At least one cave has been recognized in Chester strata just below the plane of unconformity. This cave had been filled by gray clays apparently residual from solution of the Chester limestones prior to deposition of the basal part of the Krebs strata.

The Wyandotte quadrangle is on the northwest flank of the Ozark uplift, which is a broad flat elliptical dome. The rocks for the most part are nearly flat but have a low regional northwest dip of about 20–25 feet per mile. In a few places sharply defined structural features may be accompanied by appreciable dips.

The Horse Creek anticline curves across the southern part of the quadrangle in a general northeasterly direction. It is asymmetric, with dips of 5°–18° on the southeast limb and 1° or less on the northwest limb, and thus is virtually a monocline.

The Seneca graben is a remarkably linear and persistent structural break that trends N. 40°–50° E. across the center of the quadrangle, extending both ways into adjacent quadrangles for a total length of perhaps 70 miles. It crosses the Horse Creek anticline diagonally. The structure is a combination of graben faulting and synclinal sag, having a width of 100–2,600 feet, though usually 400–1,000 feet, and a downward displacement that reaches a maximum of about 150 feet. Locally, the central block may be raised relative to the two sides and thus becomes a horst.

The Miami trough is a linear combination of syncline and graben, similar to the Seneca graben except that synclinal sag, with or without accompanying faults, prevails over true graben faulting. It crosses the western part of the Picher mining field with an average trend of N. 26° E., and is also recognizable for many miles to the northeast and southwest, over a total length of at least 42 miles. At the stratigraphic position of the ore zones in the mining field, the faults that account for part of the displacement are discontinuous and en echelon, and only in a general way do they parallel the average trend of the total structure. The width of this structure through the mining field is 300–2,000 feet, averaging about 1,000 feet. The maximum vertical displacement is about 300 feet.

The Bendelari monocline crosses the mining field with northwest trend and drops the mineral-bearing ground a maximum of 140 feet on its northeast side. The maximum dip is about 20°. Chester strata are preserved in greater thickness on the downdropped side, and the structure is hardly noticable in the Pennsylvanian strata.

The Rialto basin, involving part of the Rialto mine tract, is an irregular east-trending synclinal sump nearly a mile long, is as much as a quarter of a mile wide, and has a maximum displacement of 80 feet. It contains a thicker sequence of Chester strata than is found in surrounding ground. Other oblong or circular basins similar to, but smaller than, the Rialto basin and of no common trend occur in the mining field.

The linear structural features are of tectonic origin or have been localized on tectonic breaks, but at least the Seneca graben and Miami trough have probably been modified by intense solution of carbonate rocks in depth, with resultant subsidence. In the Rialto and smaller basins, solution and subsidence have been the dominant processes, though the solution was probably localized on deep-seated fractures or at intersections of such fractures.

Detailed contouring at the top of the Grand Falls Chert in the mining field shows a plexus of shallow basins and flat anti-

clinal ridges and domes, commonly curved, and in random orientation. Their origin is uncertain; but these features may, in part, represent initial depositional irregularities, which were also modified by compaction and settling during diagenesis.

Underground workings, particularly though not exclusively those within half a mile of the Miami trough, show numerous slump pipes in which the rocks within an irregularly cylindrical mass 100–300 feet in diameter have been faulted downward several tens of feet. The near-vertical circular fault bounding such a pipe is a poorly defined breccia zone that grades into the brecciated edges of the adjacent blocks. The block of ground within an individual pipe is generally tilted, squeezed, and mashed with some shattering of contained chert but without extreme brecciation, and clay selvages commonly develop on the bedding seams from solution of limestone beds. Slump pipes occur in areas that have been extensively mineralized, but although pre-ore, are relatively barren of mineralization. Similar slump pipes in the upper part of the Boone Formation have been mapped on the outcrop over extensive areas east and southeast of the mining field. At one place in the southern part of the Wyandotte quadrangle, Hindsville Limestone has been let down as far as the Grand Falls Chert. The slump pipes are believed to have resulted from spot solution of some very soluble underlying limestone, and they probably developed gradually as the underlying material was dissolved. Although the Joplin Member of the Boone Formation was particularly vulnerable to such solution, some slump pipes extend below this unit and others bottom above it.

Several small faults have been mapped by Branson in outcrop areas southwest of the mining field. Most of these faults trend northeast nearly parallel with the Miami trough, but two trend northwest. In the mining field, underground mapping shows, in addition to those related to the Miami trough system, scattered small faults of random orientation, some obviously related to slumpage, others not. None of these are more than 1,000 feet long, and they show a maximum displacement of 20 feet. Owing to structural incompetence of several limestone beds above the Grand Falls, probably few of these faults have much vertical extent.

Slump breccias have been produced in zones of solution, and by inference, underlie every slump pipe. They consist of insoluble materials in a residual clay matrix and have been observed underground particularly in the Joplin Member; but they also have been observed in higher members of the Boone and at the top of the Chester where they presumably underlie slump pipes that were largely eroded before the Krebs was deposited.

Tectonic breccias are abundant in the mineralized ground, particularly in the Joplin Member and in K bed of the Baxter Springs Member. They formed by plastic deformation of incompetent limestone with brecciation of the enclosed brittle chert during differential horizontal slippage between more competent beds when the strata were warped. Later mineralization has commonly replaced the limestone matrix with little or no effect on the brecciated chert.

The mineral-bearing ground was fractured along curvilinear vertical zones by joint systems along which the mineralizing solutions were introduced. These joint systems were virtually contemporaneous with the tectonic breccias. Slumpage of roof rocks owing to excess solution during mineralization produced still later fractures that are indistinguishable from the earlier joints.

The different structural features were formed at different times. Part of the structure shown in the Miami trough was developed in the interval between deposition of the Fayetteville Shale and Krebs Group, but the deformation was accentuated in post-Krebs time. The Bendelari monocline is largely a post-Fayetteville pre-Krebs structure, as are also the Rialto basin and several similar but smallers basins. The Seneca graben, on the other hand, contains enough downdropped Pennsylvanian shale to indicate that it is post-Krebs, at least in part; the Horse Creek anticline is probably also post-Krebs, though proof is lacking. Some of the slump pipes are post-Chester and pre-Krebs, but at least some of those shown in surface outcrops east and south of the mining field probably are of much later origin, forming after the Pennsylvanian shales were largely eroded.

The tectonic breccias and associated joint systems were probably produced in part by the diastrophism that preceded the Early Pennsylvanian erosion, but they were augmented by later structural movements. A Late Pennsylvanian age seems most probable to account for most of the post-Krebs deformation. There has been practically no structural deformation since the period of ore deposition which might have been Cretaceous or later.

Chert, or its cotton-rock equivalent, is present in all members of the Boone Formation. Most of the chert in structurally undisturbed ground is concordant with the bedding and has replaced limestone. This replacement is believed to have taken place on the sea floor progressively during the accumulation of the Boone, the silica being derived from the contemporary sea water. From studies in adjacent areas, the correlative part of the Mississippian was evidently a period of successive eruptions of volcanic ash, and the decomposition of this material in the sea water yielded soluble silica which accumulated in the contemporary seas and was available for replacement of susceptible sediments on the sea floor.

A little chert occurs locally as crosscutting replacement veinlets in cotton-rock outcrops and is obviously of a second generation. There is also evidence that in the period of erosion and underground solution preceding deposition of the Pennsylvanian, some silica was dissolved by circulating ground waters and reprecipitated, partly around preexisting chert masses, particularly in ground subjected to most intense leaching and condensation. This chert is not fundamentally different from the earlier generation of chert, but it is believed to be very subordinate in amount.

The total metal production of the field through 1964 has amounted to about 7,283,000 tons of zinc and 1,766,000 tons of lead. Ore minerals were first discovered by churn drilling on the southeast fringe of the Picher field about 1901, but the main part of the field was not discovered until 11 years later. Phenomental development was stimulated by the high prices for zinc during the early years of World War I, by the broad extent and shallowness of the deposits, and by the land ownership and exploitation pattern which favored the participation of a maximum number of operators. Most deposits lay between 100 and 300 feet below the prairie surface, though a maximum of 480 feet was attained in the later history of the field. Some of the richest ore in the early days showed an average mill recovery of 15 percent zinc and 10 percent lead. The peak of production was reached in 1925 when sulfide concentrates equivalent to 387,000 tons of recoverable zinc and 101,000 tons of recoverable lead were produced.

Production and grade of ore mined varied in the late 1920's and 1930's in response to economics, and many properties were shut down during the depression in the early 1930's. A secondary peak of production, only a little more than half that in 1925,

was reached in 1941, but wartime production problems and reserve depletion led to a diminishing output after that. Owing to a particularly favorable cost-price relationship, the leanest ore in the history of the field was mined during 1946 when the crude ore from the whole field averaged only 1.99 percent combined zinc and lead. During the decline of the field, economic factors tended to consolidate operations more and more under one of the major companies, the Eagle Picher Co. This company shut down operations in the summer of 1958 to await more favorable mining conditions, which are necessary to mine economically the remaing low-grade ore. Mining was resumed at a reduced rate in 1960.

The ore deposits are bodies of ore and gangue minerals that replace limestone in favorable stratigrahic zones, chiefly in the Boone Formation. Sphalerite and galena are the commercial ore minerals; sphalerite predominates. They are accompanied by a little chalcopyrite, enargite, and luzonite, but in amounts so small that the contained copper is not recovered. Marcasite and pyrite are common associates. Wurtzite is not found in the Picher field but is present at one mine near Joplin, Mo. Gangue minerals include jasperoid, dolomite, calcite, and locally, a little quartz or barite.

The sphalerite occurs in massive replacement ore and as disseminated grains, but is also abundant in vugs and other openings as crystals. These crystals are commonly a fraction of an inch to 4 or 5 inches across, or exceptionally, more than 1 foot across. The common color is rosin brown, especially on broken surfaces, but the crystals also are ruby red and black, the latter only in dolomite terrain. The crystal forms differ with the color of the crystals. Most of the sphalerite is exceptionally pure, carload lots of commercial concentrates having averaged 64.8–65.7 percent zinc, in comparison with chemically pure ZnS which contains 67.1 percent zinc. The iron content in several spectrographically analyzed samples ranges between 0.11 and 0.31 percent. The cadmium content in these same samples ranges between 0.29 and 1.4 percent, but in commercially analyzed concentrates over several years it has ranged between 0.37 and 0.56 percent. The germanium content in the several samples ranges from 0.02 to 0.10 percent, and the gallium from 0.003 to 0.060 percent. The outer black jack surfaces of crystals contain higher trace percentages of iron, copper, lead, titanium, manganese, and silver and lower percentages of cadium, gallium, and cobalt than their rosin jack cores. The ruby jack crystals show the lowest iron and the highest copper, germanium, and indium content.

The wurtzite near Joplin occurs as a minor anisotropic constituent in stalactites of sphalerite but also as small pyramidal crystals on the surfaces of these stalactites. The crystals are a new polymorph, wurtzite 10H, and a new occurrence of the polymorph, wurtzite 6H.

Galena, like sphalerite, occurs as replacement masses, disseminated grains, and crystals in open spaces. Cubic crystals are generally from a fraction of an inch to 3 inches across, but may be as much as 8 inches. Octahedral crystals are much less common, smaller, and later than the cubic crystals. Galena commonly shows pronounced etching, resolution, and resurgent crystallization, related to the primary mineralization. It carries only 0.1 ounce silver per ton.

Chalcopyrite occurs as crystals as much as ¼ inch, or rarely ¾ inch across, on top of earlier minerals in vugs and caves, less commonly as minute inclusions in sphalerite. Chalcopyrite is almost confined to dolomite-bearing ground. Enargite, chiefly in small crystals, and luzonite in smaller skeletal crystals, occur only in association with chalcopyrite. The enargite (four samples) contains 1–2 percent antimony, and 0.03–0.04 percent germanium; one analysis of the luzonite shows 1.2 percent antimony and 0.10 percent germanium.

Marcasite and pyrite are common, but the latter is in such small crystals that its total amount is insignificant. Both form crystals in openings or finely disseminated in jasperoid, and both extensively replace the earlier chalcopyrite. They also preferentially coat galena. Both minerals contain traces of nickel. Whereas the pyrite in vugs occurs in cubes or pyritohedrons or combinations of these two, the microscopic pyrite disseminated in jasperoid shows a large percentage of octahedrons.

The dolomite occurs as a massive gray spar replacement of the limestone or as pink spar crystals in veinlets and lining vugs. Calcite is a late filling of veinlets or a crystalline lining of vugs and caves. Scalenohedral crystals as much as 4 feet long that line some caves may be later than, and unrelated to, the ore-forming processes. Jasperoid is a pale-brown to blackish microcrystalline form of quartz that replaces the original limestone or gray spar dolomite. It possesses a characteristic microscopic texture in which the individual quartz grains tend to be elongate parallel to the prism axis.

The order of crystallization as established from several criteria shows numerous reversals in detail, but if only the main surges of crystallization for each mineral are considered, an order can be established into which can be fitted certain other events, as follows: Dolomite, jasperoid, sphalerite, chalcopyrite, cubic galena, marcasite and pyrite, leaching of galena (and sphalerite), resurgence of cubic galena, octahedral galena, enargite and luzonite, calcite and late chalcopyrite, flat rhombohedral calcite.

The typical ore deposit in the Picher field is confined within a stratigraphic interval that is several feet or more in thickness in flat-lying strata or gently dipping strata of the Boone Formation. In most of the ore-bearing strata, the ore bodies have a curvilinear trend following the joint systems or other channels and are called runs. There is some tendency for the runs to follow the northeast trend of the Miami trough or the northwest trend of the Bendelari monocline. In places, two or more ore-bearing zones may be superposed, and the intervening ground may be mineralized to produce a thick run of ore. Locally, Chester strata may be involved. Where the joint systems are particularly wide, the runs grade to blanket deposits having very irregular boundaries. The sheet-ground deposits in the Grand Falls Chert Member are low-grade blanket deposits, 10–12 feet thick, that have been of great commercial value because of their wide areal extent, which favors low mining costs.

The ore and gangue minerals replace the original limestone or occur as seam- and fracture-filling in the associated chert. Where structural deformation had thoroughly broken up the chert before mineralization, as in much of the Joplin Member (M bed) and in K bed, the ore is in the matrix of the chert breccia; but rich ore also occurs in these units in bordering unbrecciated ground. Here, an initial obscure layering in the limestone may be accentuated by the mineralization. Mineralized ground, whether brecciated or undisturbed, contains anastomosing replacement veinlets and pockets of gangue and ore minerals that traverse in all directions the completely replaced limestone. Vugs are ubiquitous, and ore caves are common in the limestone phases.

The sheet ground, in general, is little brecciated, and the ore has chiefly replaced the sparse thin limestone seams between the prevailing chert beds.

Within a given mineralized zone in dolomite-bearing ground, the ore and other gangue minerals tend to be zoned in relation to the dolomite periphery. Dolomite replaced the limestone in areas as much as several hundred feet across that are crudely circular or irregular in shape. These areas are surrounded by a zone in which the limestone has been replaced by jasperoid, and this zone, in turn, is bordered by one in which the limestone has been completely removed, leaving the original chert nodules and fragments as a mass of open rubble, forming what is called boulder ground. Beyond this, the ground is not mineralized, and the original limestone is unaltered and unleached, except where solution caves have developed. The greatest concentration of sphalerite is just within the edge of the dolomite, and of galena, just outside; thus, the richest stopes follow the dolomite border. The black jack form of sphalerite and the chalcopyrite, enargite, and luzonite occur only on the inner dolomite side. Although the other minerals may occur here also, they are in greater abundance on the outer wall, which is characterized by jasperoid with galena, marcasite, calcite, and some sphalerite, mostly smaller crystals of the rosin jack or ruby jack type. Ore may extend outward a short distance into the boulder ground. Beyond the limits of ore, the boulder ground may contain sparsely scattered small crystals of galena, sphalerite, marcasite, or calcite. The sequence of mineralogic zoning outward from a dolomite core is roughly the same as the order of crystallization previously cited.

Although no igneous rocks are known to which the ore deposits can be related, the ore is believed to have been deposited from hydrothermal solutions that were of magmatic origin, probably distant. Suggestive evidence for this includes the existence of a definite order of mineral crystallization with parallel zoning, comparable to that found in areas of unquestioned magmatic deposits; an indicated single episodic surge in the precipitation of each mineral, with inconsequential precipitation other than during a brief period; an experimentally indicated temperature of the solutions (115°–135°C) considerably higher than could be acquired from any surmised deep ground-water circulation, and further indicated by the "porphyroblastic" texture of the jasperoid in contrast to cold-water-deposited chert; an indicated high brine concentration in the ore solutions as determined from a study of fluid inclusions and as further indicated by the episodic leaching of galena during the mineralization; and the similarity in composition of the ores to other ore deposits whose genetic connection with intrusive igneous rocks is generally recognized.

Isotopic studies of Tri-State lead have shown that it differs from most magmatic lead in the large accretions of radiogenic lead that have taken place at some time in its past history and in the wide variation in this radiogenic accretion as between different samples. These isotopic peculiarities can be explained if the parent magma had absorbed appreciable amounts of granite from the Precambrian basement. Granite contains higher percentages of uranium and thorium and higher ratios of these elements to lead than mantle-derived rocks from which most other lead deposits have been differentiated. Hence, absorption of granite should add excess radiogenic lead from disintegration of the radioactive elements. Variation from place to place in the amount absorbed could well cause the isotopic variation of the lead. Absorption of granitic rocks by primary basaltic magma in a stable crustal environment produces alkaline rock types rich in potassium. It is precisely this type of igneous rock that is widely scattered as post-Precambrian intrusives in parts of the Mississippi Valley. In a nonorogenic environment, such rocks show only a mild tendency to intrude into higher levels, which could explain the general absence of igneous rocks in close association with Mississippi Valley ore deposits.

The age of the mineralization is not known, but certain evidence suggests that it may be later than the end of the Paleozoic.

INTRODUCTION

Location and Extent of the Field

The Picher mining field straddles the Oklahoma-Kansas State line near its east end, in Ottawa County, Okla., and Cherokee County, Kans., which are the northeast and southeast counties of these respective States. Baxter Springs, Kans., lies at the northeast corner of the field; Quapaw, Okla., lies a short distance within the southeast border; and Commerce, Okla., an early mining town, is near the southwest corner of the area that contains most of the mine workings (pl. 1). Strictly mining towns that grew up mostly in advance of the tailings piles of the mining field include Picher and Cardin, Okla., Treece, Kans., and several other named communities of indefinite boundaries, all more or less continuous with one another.

The mines of the main part of the field are included within an area that is about 9 miles long from east to west, and 8 miles broad (pl. 1). However, the workings are unevenly distributed within this area, and many of the square miles included are either barren or have made only insignificant production. Outlying mines on both sides of the State line southwest of Melrose, Kans., are 3½–5 miles west of the main field, and old workings 2 miles northwest of Miami, Okla., are about 3 miles southwest of the main field.

Regional Setting and Economic Importance

The Picher field is the westernmost of the many subdivisions, of varying productivity, that make up the Tri-State mining region. This mining region is roughly 125 miles long from west to east, and about 50 miles broad, most of it lying in Missouri. Mineralized ground is very unevenly distributed within the board limits of the region. The area near the western end has been by far the most productive. Except for minor ore occurrence at Peoria, Okla., 4 miles southeast of the Picher field, the next mining area to the east of the Picher field is at Galena, Kans.; its closest mine is about 5 miles northeast of the mine workings near Baxter Springs. The Galena mining area is connected by many small intervening deposits to the very productive mining field

at Joplin, Mo., which is similarly connected to the great mineralized field at Webb City, Mo., 24 miles northeast of Picher.

The Tri-State ores consist of dominant zinc sulfide and subordinate lead sulfide. Most mines yield both products, through some yield only zinc and a few small operations yield only lead. Ore was first discovered near Joplin in 1848. Although the lead was actively mined from that time, the associated zinc had little or no market until after the Civil War. Beginning with the rapid expansion of the zinc industry in about 1875, the Tri-State region ranked first among the zinc-producing districts of the country until very recent years, when it began to lose its dominance as a result of gradual depletion. At first the Missouri mining areas around Joplin, Webb City, and Granby were the major producers, but they were largely depleted by the end of World War I. In the meantime, parts of the Picher field had been discovered early in the present century, and the main part of the field was discovered just before World War I. Production increased rapidly during the war, and by 1917 the Picher field had assumed the leadership in zinc production, which it maintained until the 1950's. It has proved to be one of the few great mining fields of the world that have produced metal valued in excess of a billion dollars.

PURPOSE AND SCOPE OF REPORT

This report presents a detailed description of the ore deposits of the Picher field and all phases of geology having a bearing on their localization, origin, and the search for them. It is based mainly on work done by the U.S. Geological Survey, but it also incorporates pertinent data from published literature, especially from the outstanding work of George M. Fowler and associates, and some unpublished data obtained from the geologic staffs of the mining companies.

The Oklahoma part of the Picher field is in the northwestern part of the Wyandotte quadrangle. In most of the mining field, rocks at the surface consist of shale beds of Pennsylvanian age, which not only cover the ore-bearing Mississippian limestone beds but also do not reflect many of the structural features present in the ore-bearing beds. In parts of the quadrangle east and south of the field, however, the ore-bearing limestone beds and some lower strata are exposed, as are many structural features similar to those associated with the ore deposits. To show these outcrops and structures, a geologic map of the Wyandotte quadrangle, prepared for the Geological Survey in 1906–7 by C. E. Siebenthal and R. D. Mesler, is included as plate 2 of this report. The quadrangle has been made the areal unit for the stratigraphic and structural discussion in the present report.

The ore-bearing rocks and associated strata in the Picher field have been divided into many units. The variations in thickness and lithologic characteristics of these units, especially limestone grain size, chert color, and mottling, have been studied in mine workings, drill-hole cuttings, and at the outcrop. These features are described in considerable detail as they are useful in interpreting and correlating cuttings from exploratory drill holes and in understanding the localization of ore.

Although the structure of the Tri-State region generally is simple, in some places the beds have been moderately disturbed by deformation, and in other places the beds have slumped downward, probably owing to collapse into underlying solution cavities. These structures influence the vertical position of the beds and the ore deposits, and some of them may have a genetic influence on localization and origin of the deposits. For these reasons the structure of the mining area is shown in detail on maps and described in the text.

The ore deposits of the Picher field, and of the Tri-State region as a whole, are relatively simple in gross features—chiefly tabular bodies of ore minerals (sphalerite and galena) and gangue minerals (dolomite and jasperoid) that replace limestone beds; several ordinary accessory minerals are present. These minerals, however, have an interesting zonal arrangement, the crystal forms of some of the minerals change in time sequence, and in detail they have complex paragenetic relations. The minerals of the deposits are described fully because of their zonal and paragenetic distribution and genetic implications.

The deposits of the Tri-State region are representative of a general type of zinc and lead deposit that is common and economically important in many parts of the world. Characteristically, deposits of this type lack conclusive evidence of origin, and many divergent opinions have been published regarding their genesis. The present study of the Picher field has not revealed positive evidence of a specific origin; but a detailed diagnosis of the available evidence is presented, and it yields the concept of a hypogene origin. It is hoped that information presented in this report will contribute constructively to future exploration in the Tri-State region as well as to a better understanding of the origin of the deposits here and elsewhere.

GEOGRAPHY

The Picher field is on the Osage Plains, immediately adjacent to the west edge of the Ozark upland. This upland is formed on a broad low structural dome lying chiefly in southern Missouri and northern Arkansas but extending slightly into the southeast corner of Kansas and the northeast corner of Oklahoma. The dome is rudely oval in outline, the longer axis extending from southwest to northeast. Roughly it is bounded on the north by the Missouri River, on the east by the Mississippi River, on the southeast by the lower course of Black River, on the south by the Arkansas Valley, and on the west by Spring River and its extension in the lower course of Neosho River (fig. 1).

The summit of the Ozark upland along the axis of the dome is at an altitude of about 1,750 feet, some 30–odd miles east of Springfield, Mo. That part of the upland called the Springfield Plateau, which lies west from this summit, is formed on resistant cherty limestone whose average surface descends gradually westward to the general line of Spring River and the lower course of the Neosho River. A short distance beyond these rivers the top of the cherty limestone dips below the shale that forms the surface, and the Springfield Plateau there merges into the Osage Plains. Most of the Tri-State mining region lies on the general westward slope of the Springfield Plateau, but the Picher field lies in the plains area west of Spring River.

The Picher area is in general a flat prairie, treeless except for scrub timber along the courses of the shallow streams that traverse it. These streams are Tar Creek, which crosses the field from north to south, its east branch, Lytle Creek, and similar small streams lying west of the main productive part of the field, including Elm Creek, Fourmile Creek, and Squaw Creek. All flow southward to the Neosho River which lies a few miles southwest of the mining field. A short distance east of the mining field is Spring River, which is the major south-flowing tributary of the Neosho.

The streams that traverse the mining field are only slightly incised below the prairie level; but as they head only a few miles to the north, they have not proven to be particularly troublesome during periods of high water. They furnish some water for milling operations, though the chief supply is water pumped from the mines or from deep wells.

Topographic relief in the mining field is not great. The lowest point, on Tar Creek east of Commerce, is 780 feet in altitude. From here the land rises gradually to an average high of about 860 feet in the eastern part of the field, and locally, one summit is as high as 900 feet. The highest altitude in the mining field, however, is on the Blue Mound, a conspicuous sandstone hill a short distance on the Kansas side of the State line north of Picher (pl. 1). The top of this hill reaches 970 feet at one point. The maximum relief in the mining field is, therefore, about 190 feet, though most of the field is less than 80 feet above the low point on Tar Creek. Individual tailings piles from milling operations have commonly exceeded the total natural relief.

Within the Wyandotte quadrangle, the prairie extends beyond the mining field and covers most of the region southwest and west of the Neosho River, to the southwest corner of the quadrangle (pl. 2). Locally, it becomes somewhat more rolling, and may be capped by a few low sandstone hills. Southeast of Afton the prairie breaks to a slightly lower level on the southeast through some wooded hilly country along the axis of the Horse Creek anticline. A large extension of the prairie east of the Neosho River in the vicinity of Grove bears the name Cowskin Prairie; similar, but smaller, areas to the north, on the east side of the Neosho and Spring Rivers, bear the names of Swars Prairie, Jackson Prairie, and Burkhart Prairie. All these prairies east of the rivers are underlain by limestones and cherts, but they probably represent for the most part the surface from which the overlying shale has been stripped. West of the rivers, the shale still remains in the northwestern part of the quadrangle, extending south along the west edge of the quadrangle to the latitude of Afton.

Except for the prairie remnants mentioned, most of the country east of the Spring-Neosho River line is fairly well dissected by the tributaries coming into the master streams from the east. The largest of these is the Elk (or locally Cowskin) River with its north branch, Buffalo Creek. The country adjacent to the Neosho at the crossing of the Horse Creek anticline, a short distance north of the Delaware-Ottawa County line, is especially hilly, and the summits are there somewhat higher than elsewhere along the course of the river. Bluffs along the Neosho in this area were originally 250–300 feet high, and more remote summits a hundred feet higher, but Lake O' the Cherokees, made by the dam across the Neosho a short distance downstream from the southwest corner of the quadrangle, has covered the lower 60 feet of the bluffs.

Altitudes in the Wyandotte quadrangle originally ranged from about 635 feet on the Neosho at the southwest corner of the quadrangle to 1,200 feet in three widely separated places along the east boundary—the extreme southeast corner, on both sides of the head of Stogdon Hollow, and at the east end of Buckhart Prairie. Lake O' the Cherokees is now the lowest sur-

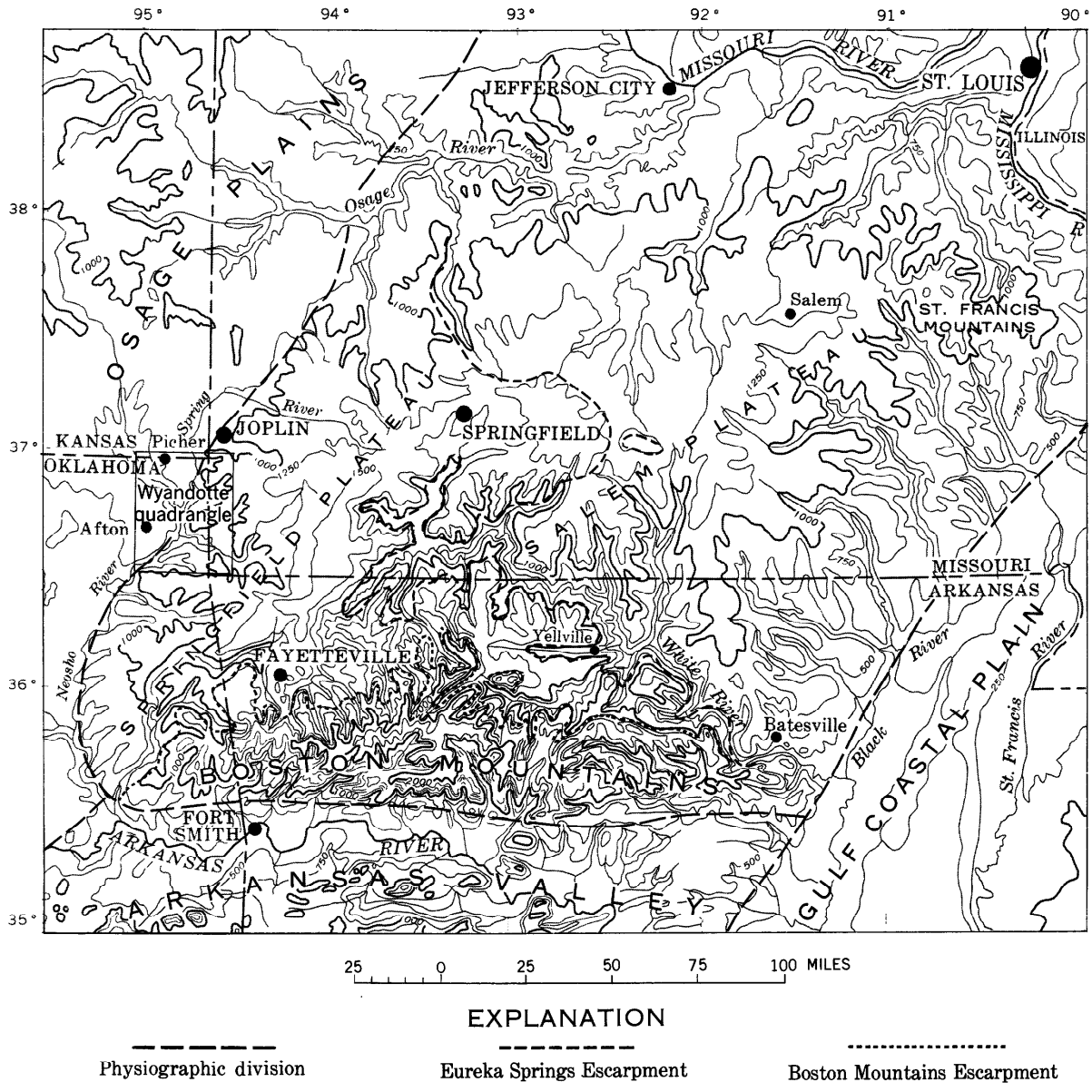

FIGURE 1.—Sketch map of the Ozark region showing its physiographic divisions and its relations to the adjacent Osage Plains, Arkansas Valley, and Gulf Coastal Plain regions. The Ozark region comprises the Boston Mountains, the Springfield Plateau, and the Salem Plateau.

face in the quadrangle, at a planned altitude of about 740 feet, though commonly maintained at the low-water stage of 735 feet.

PREVIOUS LITERATURE

The areal geology of the Wyandotte quadrangle was mapped by C. E. Siebenthal and R. D. Measler in 1906 and 1907. Siebenthal (1908) wrote a preliminary report on the mineral resources of northeastern Oklahoma, based in part on this work, and in part on reconnaissance investigation of a considerably larger surrounding area. His statement on the stratigraphy of the larger area contains references to features within the quadrangle, but the information is brief and generalized. The sketch map accompanying his report shows the locations of the Horse Creek anticline and Seneca graben

(labeled "fault"), and a good summary is given of the major structural features of the quadrangle. Indeed, the description given for the Seneca graben has been the basis for most later accounts of this feature in the geologic literature. Descriptions of the contemporary zinc-lead mines include those in the Peoria area, those along the Seneca graben, and those in the two initial discovery areas on the southeast and south fringes of the Picher field, at Lincolnville and Hattenville (Commerce).

Siebenthal's preliminary report states that the final report on the Wyandotte quadrangle was in preparation; but it was never completed. However, the geologic map compiled for that report has contributed to two issues of the State geologic map of Oklahoma, and different features of the map have been used in reports by Snider (1915), Ireland (1930), Weidman (1932), and Reed, Schoff, and Branson (1955).

In a preliminary report by Snider (1912) on the lead and zinc deposits of Oklahoma, much of the geologic information on deposits in the Wyandotte quadrangle is quoted from earlier reports on the Tri-State area, particularly the preliminary one by Siebenthal (1908). However, contemporary information on mining developments that is particularly complete for the Hattenville ("Miami") camp is given, and claim maps of the Hattenville and Lincolnville mining camps as of that date are included.

In Snider's report (1915) on the geology of northeastern Oklahoma he gave only a brief summary of the lead and zinc deposits as of that date. The major contribution of this report was on the stratigraphy of the Chester strata, based in part on statigraphic sections and collections of fossils from within the Wyandotte quadrangle.

Siebenthal's report (1915) on the origin of the zinc and lead deposits of the Joplin region covers a geographic territory far broader than the Picher mineralized field, but many features of geology in the Wyandotte quadrangle contributed to the development of the famous theory of deposition by circulating artesian ground water. The theory was evolved and written in the early stages of accelerating discovery in the main part of the Picher district, but before any extensive mine openings had been made in the field other than in the Lincolnville and Hattenville (Commerce) camps.

R. C. Moore's report (1928) on the early Mississippian formations in Missouri was a major contribution to the stratigraphy of the Ozark region, and it is as applicable to the Oklahoma counties that border on Missouri within the Wyandotte quadrangle as it is to the Missouri part of the quadrangle. Some of Moore's conclusions and correlations have been modified in a similar study of southwestern Missouri by Kaiser (1950).

No full-scale geologic account was given of the Picher district during its peak years of development in the 1920's. Spurr (1927) described certain features of the deposits based on several days of examination in the mines. Netzeband (1928) wrote a paper on the relation of fracture zones to ore bodies, and two papers (1929a, b) on specific mine development subjects, which also gave brief but good descriptions of the deposits and their stratigraphic occurrences.

A brief account of the geology in Ottawa County is given by Ireland (1930) in a report on the prospects for oil and gas in three northeast Oklahoma counties. Other papers that appeared during the 1930's which have varying relevance to the stratigraphy of the area were by Cline (1934), Laudon (1939), and Pierce and Courtier (1938), the latter treating the geology of the coal fields in Cherokee County and two adjacent counties in southeast Kansas.

Weidman (1932) described the geology and ore deposits of the Oklahoma part of the Picher field. The geologic map of Ottawa County is stated to be based on maps by Siebenthal, but there were extensive revisions in detail, or perhaps in part generalizations, and new mapping was added showing the outcrop of sandstone members in the Pennsylvanian strata. About one-fourth of the Oklahoma mines were described in detail; these descriptions furnished the basis for discussion of the geologic features and conclusions concerning the ore deposits.

In 1932 also appeared the first of an important series of papers by Fowler and Lyden, and by Fowler and others (see next paragraph) describing the geology of the Tri-State area with particular reference to the Picher field. The first clear-cut subdivision of the Mississippian strata into recognizable stratigraphic units was made. These units in the Boone Formation were designated, from the top down, by capital letters from B to R, and the varying susceptibility of the different units to mineralization was outlined. The designations have been widely accepted and used by geologists, engineers, and operators in the field to great profit in prospecting and developing the ore bodies. The relationship of the ore bodies to tectonic deformation was particularly emphasized, and the concept of secondary origin of the abundant chert in the Boone Formation was also stressed.

Later papers (Fowler, 1933; Fowler and Lyden, 1934; Fowler, Lyden, Gregory, and Agar, 1935; Fowler, 1938, 1942, 1943; Fowler, Hernon, Conrow, and Stone, 1955)

elaborated or modified the conclusions of the 1932 paper by Fowler and Lyden, or presented additional facets of the geology.

A symposium by Bastin and others (1939) on the lead and zinc deposits of the Mississippi Valley region includes much information pertinent to the Picher field. This report details the stratigraphic and structural setting, the mineralogy, and any association of igneous rocks in the various districts, and ends with a general discussion of origin. Additional papers dealing with origin were written by Emmons (1929), Ridge (1936), and Garrels (1941). All these papers support to varying degree the theory of deposition from hydrothermal solutions of igneous origin.

The results of various types of geophysical investigation in the Tri-State district were given in a report by Jakosky, Dreyer, and Wilson in 1942; it was supplemented by a later study of radioactivity (Dreyer, 1948). Stoiber (1946), from a study of asymmetric crystal growths, mineral overgrowths, and the position of crystals on cavity walls in the Picher deposits, concluded that the depositing solutions followed the regional facture pattern and moved outward from the major structural break crossing the field, the Miami trough. McKnight (1948), in a general summary on the deposits of the Tri-State region, described the common occurrence of ore runs around a dolomite core, with associated mineralogic zoning. Lyden (1950) reported on the geology of the Picher field, with particular reference to those features that could be used as guides in ore finding. In addition to structural features previously used in the field, he demonstrated the usefulness of maps showing the relation of the ore deposits to the distribution of dolomite and jasperoid. This paper also described the peculiar structural features called pipe-slumps (or slump pipes) and pointed out the relation of ore deposits to them.

In 1955, Reed, Schoff, and Branson reported on the ground-water resources of Ottawa County. Much new information is given on the subsurface geology as obtained from a study of well cuttings, and particularly from insoluble residues from such cuttings. Description of the Mississippian formations is largely abstracted from the literature, but the chapter on the Pennsylvanian, by Branson, subdivides the strata of this period into several units which are described and correlated with the standard Oklahoma section in the McAlester basin. The new units are shown on the geologic map accompanying the report. However, a report by Howe (1956), treating the same sequence of rocks across the State line in Cherokee County, Kans., gives an entirely different stratigraphic classification.

Harris (1956) included localities in Ottawa County in his study of Chester strata in northeastern Oklahoma. Extensive fieldwork in northeastern Oklahoma in preparation for a new State geologic map (Miser, 1954) was coordinated and summarized in a general report on the geology of the flanks of the Ozark uplift by Huffman (1958).

PRESENT INVESTIGATION

As the geologic report on the Wyandotte quadrangle contemplated by Siebenthal was never completed, the present investigation may be considered as having started in 1906. Most of the areal geologic mapping in the quadrangle was completed by Siebenthal and Mesler in 1906–7. Development in the Picher mining field was accelerated shortly thereafter, and the attempt was evidently made by Siebenthal to examine all mine workings in order to prepare a complete report on the mineral resources of the quadrangle. Few of his notes are dated; but apparently some fieldwork was done in 1910, and extensive field investigations were carried out in 1913. Assumption of responsibility in 1907 for compilation of resource data on lead and zinc in the midcontinent area, and assignment to other related duties during World War I, interferred with the progress of the fieldwork, and the phenomenal expansion within the Picher field during the war soon made hopeless the task of trying to keep abreast of the underground developments. Considerable underground mapping was done after the war, but poor health and increasing preoccupation with other duties prevented completion of that task. Siebenthal died in 1930.

A short generalized text on the Boone and earlier strata was left by Siebenthal, but nothing on later formations. Because the stratigraphy within the Boone had not been worked out, this manuscript was obsolete. However, the field notebooks of both Siebenthal and Mesler, covering the areal mapping, have been used extensively in interpreting the geology of the quadrangle in the light of later information. Their geologic map is herewith published for the first time in its entire detail. Except that the Boone was incompletely subdivided, the map is adequate for the present study.

A new project was started in the mining field in March 1934, with E. T. McKnight in charge. The work consisted primarily of underground mapping of the ore deposits on a scale of 100 feet to the inch; this was supplemented by a study of drill cuttings within the mining field and of surface exposures at favorable outcrops within the Wyandotte quadrangle. No attempt was made to remap the areal geology of the quadrangle, though a few additions or corrections have been made.

Geologists associated directly on the project included Carl C. Addison to the end of 1935, Kenneth R. Bowie and Joseph M. Thiel to mid-1935, Francis G. Wells for a few weeks in the winter of 1934–35, Irving T. Schwade for a few months in 1936, and R. P. Fischer from October 1937 until the termination of the fieldwork in the spring of 1941. At different times M. F. Owens, Jr., Donald Bradley, Dee T. Waters, Jr., and Roy Brown assisted in the fieldwork.

In March 1939, McKnight returned to Washington. Fischer stayed on in the district for a few weeks to continue compilation of the structure map, and after spending the summer on another assignment, he returned to the district in the fall to spend the ensuing winter office season working on the structure compilation. In this compilation, the information obtained from our underground mapping and from study of well cuttings was supplemented by well-log information obtained from the offices of the various mining companies. The job was completed in the spring of 1941.

During April 1940, Fischer mapped in detail the geology of that part of the Neosho basin which was being flooded by Lake O' the Cherokees. This mapping was carried on just ahead of rising water behind the closed watergates of the dam. Stratigraphic information that was obtained coincident with the mapping has proven invaluable in rounding out the regional stratigraphy.

To supplement and strengthen the stratigraphic phases of the project, James Steele Williams, of the Geological Survey, spent 4 weeks in May and September 1934 and 5 weeks in October and November 1936 measuring stratigraphic sections and collecting fossils from available exposures of Mississippian and Pennsylvanian rocks within the Wyandotte quadrangle. He was accompanied for 6 days in October 1936 by George H. Girty, of the Geological Survey, who had, many years earlier, visted and collected fossils in the field with Siebenthal, and had received additional collections made by Siebenthal and Paul V. Roundy. Williams turned over his collections and notes, as well as those of Girty, to Mackenzie Gordon, Jr., of the Geological Survey, who assumes responsibility for the stratigraphic correlations of Mississippian formations made in the present report. Gordon has visited and collected from various localities in the Wyandotte quadrangle at several times since September 1941.

Because the mine workings in the Picher field were more extensive than could be mapped in the time allotted to the underground work, from the beginning of the new project in 1934 the work was carried out in a rough checkerboard pattern, so that deposits in most parts of the field were mapped to varying degree. A large block that was not mapped lay in Kansas north of the Robinson and Fox mines, but there were also large gaps in the eastern part of the field where most of the mines were inaccessible during the period of our fieldwork. The geologic map of the field is, therefore, incomplete. Enough mines in the main part of the field were mapped to give a fairly continuous block east to the Dardene, Maxine, and Pat tracts (pl. 1), and a preliminary map showing the geologic structure and distribution of dolomite in this block was published in 1944 (McKnight and others, 1944). No attempt has been made to compile a map from the information available on the scattered deposits mapped to the east of this block, as the areas mapped are meager in comparison to the total area. The Eagle Picher Co. very kindly consented to furnish available structural information to help fill in the gaps in our published preliminary map. McKnight, therefore, spent a month in Cardin, Okla., during March and April 1955, compiling additions to the structural map. Some of the Eagle Picher data were available in structure maps on the same stratigraphic datum as our compilation, and some were on a different datum that had to be adjusted to ours. These additions are based on geologic mapping or stratigraphic logging of well cuttings by the following present or past members of the Eagle Picher geologic staff: Douglas C. Brockie, Harry M. Callaway, Norman E. Eastmoore, Jr., Perry K. Hurlbut, Andrew Kuklis, Joseph P. Lyden, and Curtis Templain.

Completion of the present report has been long delayed by assignment of the senior author to other duties from well before World War II until after the Korean War.

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STRATIGRAPHY

General Features

Within the Picher mining field, the rocks exposed at the surface encompass a relatively thin interval of the Mississippian and Middle Pennsylvanian parts of the geologic column. Thinness of this interval is the result of two independent factors, namely, the low dip of the strata and the flatness of the prairie surface on which they crop out. Stratigraphically lower sections of the Mississippian are cut in the mine workings and prospect drill holes.

Over the larger area included within the Wyandotte quadrangle, more pronounced structural features and a greater topographic relief combine to expose a somewhat thicker section, including Middle and Lower Pennsylvanian, all the Mississippian, the Upper Devonian, and the upper part of the next underlying unit which is of Early Ordovician age. The maximum thickness exposed is about 1,000 feet (fig. 2), though this is a composite thickness and only part is exposed at any one place. Major unconformities occur at the base of the Upper Devonian and at the base of the Middle Pennsylvanian.

The exposed section is predominantly limestone and chert, with some dolomite, shale, and sandstone, beneath a capping of Pennsylvanian black shale. The lower shales, which comprise the Upper Devonian, one formation of the Mississippian, and parts of a Lower Pennsylvanian formation, crop out in the southern half of the Wyandotte quadrangle; but they are thin or missing in the subsurface section of the mining field, chiefly owing to the unconformity at the base of the capping Pennsylvanian shale.

Deep water wells and prospect drill holes in and adjacent to the mining field have cut several hundred feet of strata below the exposed section. These are of Early Ordovician and Late Cambrian age and consist predominantly of dolomite with some sandstone and sparse shale; their closest outcrops lie many miles to the east on the flanks and top of the Ozark dome in Missouri. Nine of the holes have cut through these lower Paleozoic strata and penetrated the granitic basement rocks which are of Precambrian age. The maximum thickness of the Cambrian and Ordovician revealed is about 1,325 feet, of which only the top 26 feet is represented in outcrops in the quadrangle. Two additional prospect holes have gone directly from Mississippian strata into buried peaks of the Precambrian terrain.

System	Series	Group, formation or member	Columnar section	Thickness (feet)	Description
PENNSYLVANIAN	Des Moines	Krebs Group: Bluejacket Sandstone Member (of Boggy Formation)		15-60	Brown to buff sandstone.
		Savannah Shale / Doneley Limestone Member of Branson, 1954		120±	Black and gray fissile shale, a little sandstone, thin black fossiliferous limestone (Doneley Member), thin coal and underclay seams (Branson, 1955).
		McAlester Shale / Warner Sandstone Member		30± / 0-20	Black fissile shale with clay ironstone concretions, sparse siltstone, thin coal and underclay; brown coarse-grained sandstone (Warner) at base (Branson, 1955).
		Hartshorne Formation		0-50	Dark-gray to black fissile shale, subordinate siltstone, sparse calcareous clay ironstone, and thin coal seams with underclay (Branson, 1955).
		UNCONFORMITY			
	Morrow	Hale Formation		0-83+	Alternating brown to black carbonaceous and locally ferruginous sandstone, dark shale, and fossiliferous bituminous limestone, partly oolitic.
		UNCONFORMITY			
MISSISSIPPIAN	Chester	Fayetteville Shale		0-70	Black, bluish-gray, and greenish fissile or limy shale with local ironstone concretions, subordinate gray and brown to purplish crinoidal limestone, part bituminous, part oolitic.
		Batesville Sandstone		0-70	Gray crinoidal to dense limestone, commonly oolitic, buff sandstone and green shale, interbedded.
		Hindsville Limestone		0-85 ±	Gray crinoidal to dense limestone, commonly oolitic, locally cherty, a little sandstone and green shale.
		UNCONFORMITY			
	Meramec	Quapaw Limestone		0-31+	Gray medium- to coarse-grained crinoidal limestone.
		Boone Formation: Moccasin Bend Member		0-140	Alternating chert and fine- to medium-grained brown limestone, some cotton rock; chert conspicuously brown and blue in lower part, paler above.
		Baxter Springs Member		0-5 / 0-51 / 0-60	At base, bedded to massive pale chert or cotton rock, glauconitic at base (L bed); overlain and overlapped regionally by crinoidal glauconitic limestone and variegated chert, the limestone locally shaly or containing glauconitic oolite and phosphate nodules (K bed); topped by thin phosphatic and highly glauconitic crinoidal limestone containing variegated and, in part, very dark chert (J bed).
		Short Creek Oolite Member		0-10	Brown oolitic limestone, only slightly glauconitic.
		DISCONFORMITY			
	Osage	Boone Formation: Joplin Member		0-100	Gray crinoidal limestone and nodular or bedded chert; chert-free ledge near base.
		Grand Falls Chert Member		25-95	Pale chert, cotton rock, and subordinate brown fine-grained limestone.
		Reeds Spring Member		70-105	Blue, gray, and brown chert alternating with gray and brown fine-grained limestone; crinoidal bioherms locally at base.
		St Joe Limestone Member		10-32	Gray to pink crinoidal limestone with massive ledge at top and greenish shaly zone below middle; sparse blue to gray chert.
MISSISSIPPIAN AND DEVONIAN	Kinderhook and Upper Devonian	Chattanooga Shale		0-50	Black fissile shale, bleached greenish or yellow at top; locally a few inches of coarse-grained white sandstone at base.
		UNCONFORMITY			
ORDOVICIAN	Lower Ordovician	Cotter Dolomite		26+	Gray to brown dolomite, fine-to medium-grained, locally sandy; a little chert, in part oolitic.

FIGURE 2.—Generalized section of rocks exposed in the Wyandotte quadrangle.

PRECAMBRIAN ROCKS

In 1962–63, the Eagle Picher Co. drilled six deep prospect holes to the Precambrian basement in the Picher field. Two earlier churn-drill holes—the Bird Dog and the John Beaver deep wells—had also reached the Precambrian. A zigzagging section (fig. 3) shows the depth to the Precambrian in these eight holes. We are indebted to the Eagle Picher Co. for the information on the new drilling and for permission to publish this information.

Depth to the Precambrian ranges from 1,246 feet (about 425 ft below sea level) in the Bird Dog deep well to 1,944 feet (1,085 ft below sea level) in the Kansas Ex. prospect hole P54, 2¾ miles away. The latter hole was drilled in the Miami trough, a narrow graben block having a displacement of about 300 feet in the vicinity of the drill hole. Except in this graben block, the deepest Precambrian is at 1,770 feet (about 937 ft below sea level) in the John Beaver well.

B. F. Goodrich well 3 at Miami (sec. 24, T. 28 N., R. 22 E.) hit the basement granite at 1,045 feet (about 230±15 ft below sea level). Two other deep wells on the Goodrich tract that are collared at nearly the same altitude and less than a quarter of a mile to the west and to the north of well 3 failed to hit granite at depths about 155 and 420 feet below the level of the granite contact in well 3 (Reed and others, 1955, p. 36), showing that the local relief on the granite surface is pronounced.

Another area where the Precambrian granite has been reached by drilling lies on the Demo tract west of Commerce and a short distance east of the Neosho River. The granite here lies at surprisingly shallow depth. In 1959 the American Zinc Co. deepened two holes, about 300 feet apart, in the NE¼SE¼ sec. 8, T. 28 N., R. 22 E. (pl. 2). The west hole, according to logging by E. H. Hare and R. A. Chadwick, of the Eagle Picher Co., reached red porphyritic granite at a depth of about 291 feet (495±10 ft above sea level), and the east hole reached it at a depth of 345 feet. But a drill hole starting at nearly the same altitude on the flat prairie 1,100 feet to the northeast penetrated 650 feet without reaching the granite.

In a discussion of the relief on the granite surface in the Picher-Miami area, Schoff (Reed and others, 1955, p. 36) states that "further evidence indicating an irregular surface on the granite is the absence of some of the lower sedimentary strata in the vicinity of the granite 'peaks'—as if the peaks had been islands in the sea while the oldest sediments were being deposited." The granite peak on the Demo tract is in contact with stratigraphic units in the Boone Formation (Mississippian); [1] thus, the lower part of the Boone and about 1,300 feet of Ordovician and Cambrian strata, present elsewhere, are missing. The relations are comparable to those at the outcrop area in the St. Francois Mountains on top of the Ozark dome in southeastern Missouri. There, the local relief on the landmass of Precambrian rocks that was submerged at the beginning of the Paleozoic sedimentation approximated 2,000 feet (Dake, 1930, p. 194), and younger Paleozoic formations overlapped older ones to lie in contact with Precambrian rocks high on the flanks of the buried hills (Dake, 1930, geologic map).

The Precambrian cut in the deep drill holes has generally been described in logs as "granite," or "pink granite." However, five of the six recent Eagle Picher holes in the Picher field intersected fine-grained rock types, and a preliminary examination of this material suggests that it may be from igneous flows (Douglas C. Brockie, written commun., Feb. 11, 1964).

The basement granite in the mining field has been interpreted by some geologists as intrusive into the Paleozoic strata. The same interpretation has been applied to the granite that crops out along a narrow belt in the valley of Spavinaw Creek a few miles southwest of the Wyandotte quadrangle. However, the lack of thermal metamorphism both at Picher and Spavinaw Creek, the occurrence of detrital granite boulders and feldspar grains in the dolomite bordering the granite at Spavinaw, and the analogy with outcrop areas in Missouri are believed to be convincing for a Precambrian age (see Tolman and Landes, in Bastin, 1939, p. 76–81; Ham and Dott, 1943).

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[1] Interpretation of the stratigraphic section cut by the two drill holes above the granite peak has to be based mainly on driller's logs of the initial holes, before deepening. The log of the west hole is particularly hard to interpret, but apparently the hole went from the Joplin Member of the Boone Formation (M bed of Fowler and Lyden) into the granite. The east hole, which hits the granite at a level about 54 feet lower, shows what is believed to be the Reeds Spring Member of the Boone in contact with the granite.

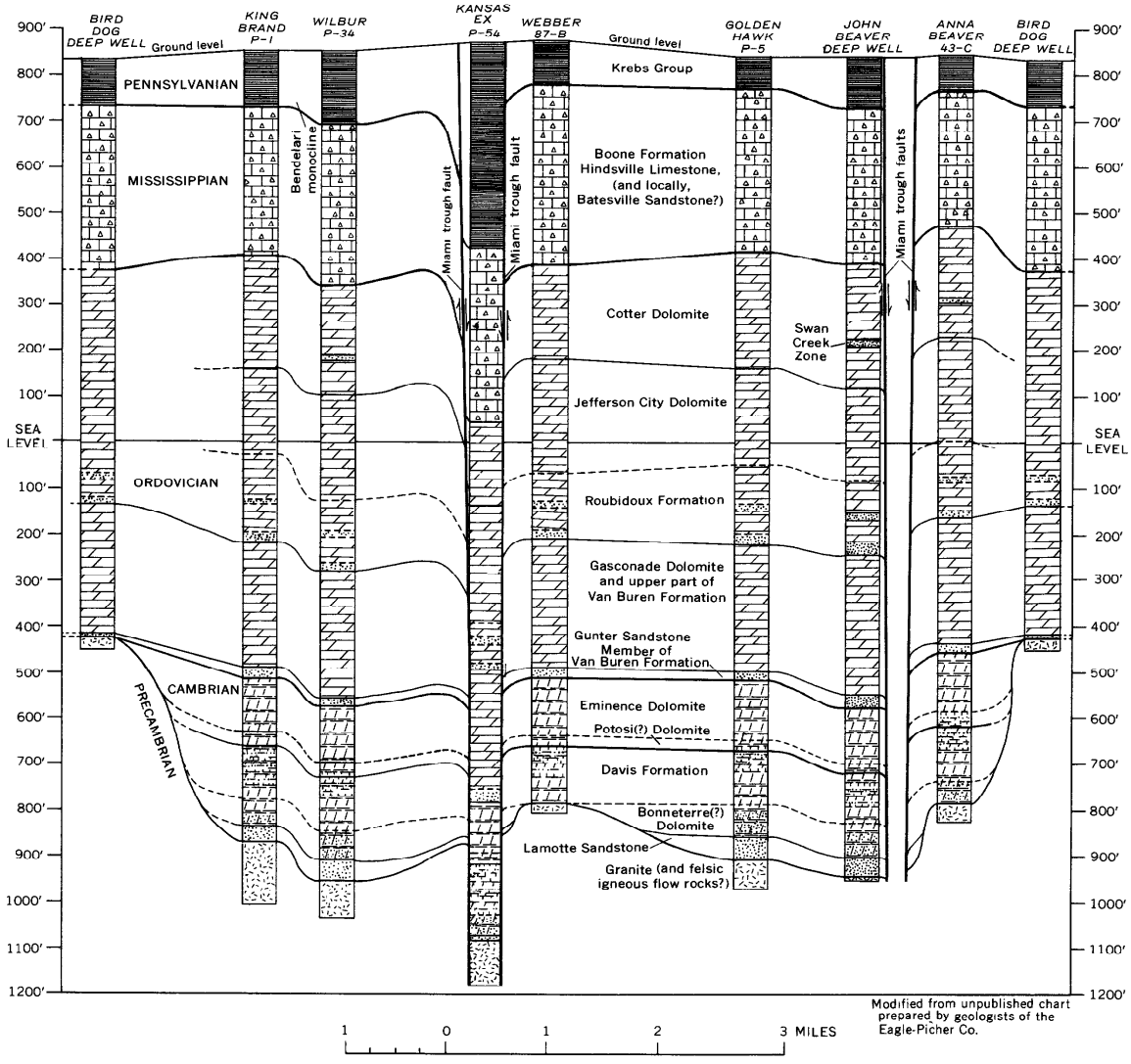

FIGURE 3.—Cross section through deep drill holes that reached Precambrian rocks in the Picher field.

CAMBRIAN AND ORDOVICIAN SYSTEMS

UPPER CAMBRIAN AND LOWER ORDOVICIAN SERIES

Lower Paleozoic formations are important sources of domestic and industrial water in the Tri-State region, and they have been studied intensively where they crop out in Missouri and in other places where they have been cut in deep wells. These formations in northeastern Oklahoma have been described in detail by Reed, Schoff, and Branson (1955), in connection with ground-water studies, and much of the following summary is abstracted from their report, though modified by information from the recent Eagle Picher deep drilling.

The lower Paleozoic formations of the Ozark region consist dominantly of dolomite, with some chert, sandstone, and minor shale. Because the dolomite of the different formations looks much the same in well cuttings, a technique has been perfected by geologists of the Missouri Geological Survey wherein the cuttings are first leached with acid to destroy the dolomite, and criteria distinctive for the different stratigraphic units are recognized through a study of the insoluble residues. The stratigraphic classification of Cambrian and Ordovician strata given in the ground-water report cited has been made, in part, by geologists of the Missouri Survey from a study of cuttings from 12 Oklahoma wells, and, in part, by Ireland (1944) from a study of the early Paleozoic stratigraphy of northeastern Oklahoma by means of insoluble residues.

Table 1 gives the Cambrian and Ordovician formations, their lithologic character, and thicknesses, with the variation in thickness indicated. The lowermost formations are encountered only in the exceptionally deep drill holes; hence, not so many intersections have been made of them as of the higher formations. As the structural dip off the Ozark dome is nearly flat, the formational intercepts encountered in the vertical wells can, in general, be considered as stratigraphic thicknesses.

All the formations given in table 1 were cut in five of the deep holes drilled in the Picher field (fig. 3); but in other holes which hit the Precambrian at shallower depth, the lower formations are missing. Thus, B. F. Goodrich well 3 went from Roubidoux Formation into the granite, indicating that the Roubidoux had stratigraphically overlapped all earlier formations to rest against a buried granite ridge or peak. On the Demo tract all Cambrian and Ordovician strata are pinched out against the buried granite hill.

Because the top part of the Cotter Dolomite crops out in the Wyandotte quadrangle, this formation is discussed a little more fully in the following pages.

TABLE 1.—*Lower Paleozoic formations in deep wells and prospect drill holes, Picher mining field and vicinity*

[Modified from Reed and others, 1955; Ireland, 1944; unpub. data from Eagle Picher Co., 1963]

Series	Formation or member	Lithologic character	Thickness (feet)
Lower Ordovician	Cotter Dolomite (including Swan Creek zone at base).[1]	Dolomite, a little of it sandy, particularly at base; some chert which is in part oolitic or sandy; a little shale.	143–183 [1] (16 holes).
	Jefferson City Dolomite [2]	Dolomite and chert, in part oolitic.	270–340 (15 holes).
	Roubidoux Formation	Dolomite and chert, in part oolitic; 2 or 3 layers of sandstone	142–190 (14 holes). 105 (1 hole).
	Gasconade Dolomite and upper part of the Van Buren Formation.	Dolomite and chert	240–300 (10 holes).
	Gunter Sandstone Member of Van Buren Formation.	Sandy dolomite	20–40 (9 holes).
Upper Cambrian	Eminence Dolomite (possibly including thin Potosi Dolomite at base).	Dolomite and shaly siltstone with some chert, oolite, and glauconite; pyritic.	135–157 (7 holes).
	Davis Formation	Dolomite, magnesian limestone, and shaly siltstone, with some sandstone, chert, and glauconitic shale; locally pyritic.	110–120 (7 holes).
	Bonneterre(?) Dolomite	Sandy dolomite and shaly siltstone, with some sandstone and shale; glauconitic, locally pyritic.	45–75 (5 holes.)
	Lamotte Sandstone	White sandstone, detrital igneous material.	12–50 (5 holes).
Precambrian		Granite or salic(?) igneous flows.	(11 holes).

[1] The thickness of Cotter Dolomite herein cited follows the interpretation of Ireland (1944), who considers the Swan Creek zone as the base of the Cotter. Earl McCracken, Missouri Geological Survey, placed the base of the Cotter somewhat lower in the section with corresponding restriction of the underlying Jefferson City, and this interpretation is the one followed in unpublished material from the Eagle Picher Co., reproduced in fig. 3, and followed also by Branson (1944, p. 55–56). By this interpretation the Cotter in six of the holes in relatively undisturbed ground in the Picher field (fig. 3) ranges from 204 to 268 ft, four of them from 238 to 251 ft. According to McCracken (written commun., 1956), thickness of the Cotter in the Ballard mine deep well near Baxter Springs is 220 ft (above 205 ft of Jefferson City) with the Swan Creek zone in the interval 140–145 ft below the top of the Cotter.

[2] The Theodosia and Rich Fountain Formations of Cullison (1944), which occupy the interval included here in the Jefferson City and possibly some of the lower Cotter, have not been distinguished in the stratigraphic terminology used by those who have studied insoluble residues from Oklahoma localities.

ORDOVICIAN SYSTEM

LOWER ORDOVICIAN SERIES

COTTER DOLOMITE

The Cotter Dolomite was formerly exposed in two localities at the foot of bluffs along the Neosho River—one over a stretch of almost 2 miles on the left bank of the river between 1 and 2 miles north of the Delaware-Ottawa County line, and the other over a stretch of about half a mile on the left bank between the mouths of Honey Creek and Woodward Hollow, southwest of Grove (pl. 2). Both localities are now beneath the level of Lake O' the Cherokees. The formation is exposed where the Horse Creek anticline crosses Buffalo Creek, 3 miles northeast of Tiff City, and also in the bed of the Elk River just east of the quadrangle boundary. It has been cut in the mining field only in a few deep prospect holes or wells drilled for water in underlying sandstones.

THICKNESS

At the locality along the Neosho River north of the Delaware County line, the highest exposure of the Cotter was about 18 feet above the former low-water stage of the river. The outcrops were poor, and a maximum of only 11 feet at the top of the formation was exposed. On Buffalo Creek, 26 feet of the formation is exposed above the creek level, and at the Elk River locality, about 12 feet. In deep drill holes in and adjacent to the mining field, thicknesses based on a study of insoluble residues range from 143 to 183 feet, according to one set of determinations (Reed and others, 1955, p. 44), but another interpretation on the stratigraphic extent of the Cotter takes in additional strata at the base, to give a total thickness of 204–268 feet (fig. 3; table 1).

CHARACTER

The dolomite may be fine grained, thin bedded, or medium grained, and thin bedded to somewhat more massively bedded. It is gray to brownish gray or locally brown and weathers pinkish tan to yellowish gray to pale gray. Rounded sand grains are imbedded in part of the dolomite in the surface exposures; they may be massed in certain thin seams, 1–6 inches thick, and the intervening dolomite may not be noticeably sandy. But no appreciable sandy phase is present in the section ascribed to the Cotter in the Ballard deep well. A little chert is scattered througout the formation; it is usually white or pale gray, but in part, gray, brown, light blue, and dark blue. The more opaque cherts are commonly oolitic. Where seen in surface exposures of the upper part of the Cotter, the chert is in lenses or irregularly rounded masses about 2 inches in diameter. However, the cherts cut in the Ballard well may be partly bedded, but if so, the beds cannot be more than a 1–2 feet thick.

CORRELATION

No fossils have been found in the Cotter in the Wyandotte quadrangle. The designation as Cotter is based in part on stratigraphic correlation of deep well cuttings made by Ireland (1944) and by members of the Missouri Geological Survey (see Reed and others, 1955, p. 143), using the insoluble residue technique, and in part on lithologic similarity to strata that have yielded Cotter fossils in exposures a short distance south of the quadrangle.

The Cotter has been traced from its outcrops in central southern Missouri through subsurface well cuttings to Carthage, a few miles northeast of Joplin, and has thicknesses of 270–350 feet lying above 50–70 feet of Jefferson City Dolomite in the several wells studied (McQueen, 1931, pl. 14). A study of insoluble residues from deep wells in and adjacent to the Picher mining field shows 143–183 feet of Cotter, including at its base 15–40 feet of the Swan Creek, lying above 270–340 feet of Jefferson City (Reed and others, 1955, p. 43–44; Ireland, 1944). But this interpretation within the Picher field places the base of the Cotter higher in the stratigraphic section than does another interpretation that has been followed by some investigators (see table 1, footnote 1). If additional strata below the Swan Creek are assigned to the Cotter, in accordance with this second interpretation, thicknesses in the Picher field are more nearly in accord with those cited in Missouri, and indicate that the correlation from insoluble residues is correct.

On Spavinaw Creek, about 10 miles south of the southwest corner of the Wyandotte quadrangle, 125 feet of dolomite is exposed in the same stratigraphic position as the dolomite that formerly cropped out along the Neosho. Gore (1952) assigns these strata to the Cotter Dolomite, basing the correlation on fossils collected by others. A composite collection from these strata, made by several geologists and deposited in the U.S. National Museum, was derived almost entirely from the immediate vicinity of the Precambrian granite outcrops about half a mile southwest of Spavinaw (H. D. Miser, oral commun., 1967). The stratigraphic horizon at this locality, according to the information published by Gore (1952), is near the base of the exposed section. The fossils have been examined by Ellis Yochelson, of the U.S. Geological Survey, who expresses the opinion

(written commun., 1966) that the stratigraphic interval represented is included in the unit for which Cullison (1944) introduced the term Theodosia Formation. This unit is the equivalent of lower Cotter or upper Jefferson City (Cullison, 1944, p. 25).

On the basis of a well-known, widely occurring fauna in Missouri and northern Arkansas, the Cotter is classed as Lower Ordovician (Ulrich, in Purdue and Miser, 1916, p. 4).

Stratigraphic Relations

The Cotter Dolomite, according to Branson (1944, p. 56), is "uncomformable upon the underlying Jefferson City formation, although there is but slight field evidence of pre-Cotter erosion." Cullison (1944) described the Theodosia Formation, to include strata that had been previously called "upper Jefferson City" and "lower Cotter Dolomite" by various workers. He studied and was the first to designate a type section at the type locality of the Cotter Dolomite. The Cotter, as defined by him, is unconformable on the Theodosia, although the hiatus between the two formations was relatively short, and there is no evidence that the two formations are other than parallel.

The Cotter is overlain, with pronounced unconformity though without angularity, by the Chattanooga Shale whose basal part is of Late Devonian age.

Devonian and Mississippian Systems

Upper Devonian and Kinderhook Series

Chattanooga Shale

The Chattanooga Shale is exposed along the Elk River at a few localities in the Missouri part of the Wyandotte quadrangle, and also where the Horse Creek anticline crosses Buffalo Creek 3 miles northeast of Tiff City. It was formerly exposed near the foot of the bluff and in adjacent hollows along the left side of the Neosho River, between 1 and 2½ miles north of the Delaware-Ottawa County line, but now only 8 or 10 feet at the top of the formation crops out in the structurally highest parts of this area at the low-water stage of Lake O' the Cherokees (735 ft alt). Former outcrops at several localities along the river and in two tributary hollows southwest of Grove are now covered by the lake (pl. 2).

Thickness

The shale is 63–67 feet thick in four test holes drilled on the west side of the Neosho River at a rejected dam-site crossing the township line just south of Tynon Bluffs. It is 32 feet thick in former exposures, now under water, at the mouth of Honey Creek, southwest of Grove; 50 feet is reported in a well at South West City, and 28 feet in a well just 1 mile east of Tiff City. It is 20 feet thick where the Horse Creek anticline crosses Buffalo Creek, 26–34 feet thick at submerged exposures where the Horse Creek anticline crosses the Neosho River north of the Delaware County line, 25 feet thick in the city well at Fairland (Weidman, 1932, p. 11), and 14 feet thick in the city well at Miami (Weidman, 1932, p. 11).

Weidman (1932, p. 11) cites several thicknesses from deep wells in the vicinity of Miami, and three from deep mine wells in the Picher field. These thicknesses show a decrease from a maximum of 15 feet near Miami to a thickness of 1–5 feet in the mining field. However, the occurrences in the mining field (Rialto, Lucky Syndicate, and Victory Metal Co. mines) appear to be exceptional for the logs of numerous other deep wells on widely scattered tracts, including two on the Eagle Picher Central Mill tract(NW¼ sec. 31, T. 29 N., R. 23 E.), do not record any black shale (Reed and others, 1955, app. A, B). The shale is also absent in deep holes at the Barr and Ballard mines, just across the State line in Kansas. The most northerly consistent occurrence of the shale is on the B. F. Goodrich and McCoy greenhouse tracts in the north outskirts of Miami, where the thickness ranges from 5 to 8 feet (Reed and others, 1955, app. A).

Character

The Chattanooga is typically a fissile black shale, generally poorly exposed except at the top where it is protected from weathering by the limestone ledge at the base of the Mississippian. At former exposures below the mouth of Honey Creek, the top 4½ feet is poorly bedded, nonfissile gray to dark-gray shale that is sandy in the basal foot of this interval. The shale there becomes greenish gray in the top 6–12 inches. A similar bleaching to greenish gray or yellow in a thin layer near or at the top has been noted in other places.

In the exposures on Buffalo Creek, 1–4 inches of coarse, somewhat quartzitic, white sandstone, containing numerous brown spots presumably from the weathering of pyrite, lies at the base of the shale, in the position occupied by the Sylamore Sandstone Member in neighboring parts of Missouri and Arkansas. This sandstone has not been observed in other outcrops in the quadrangle, but may be present, without the overlying shale, in the buried section in the Picher mining field.

Correlation and Age

The black shale is an easily recognizable unit between the Ordovician dolomite below and Mississippian lime-

stone above. Although it is not continuously exposed over the southwest flank of the Ozarks, the gaps between continuous exposures are small enough to assure that the unit in the Wyandotte quadrangle is the same as that widely mapped as Chattanooga in northeast Oklahoma, northwest Arkansas, and southwest Missouri. No fossils have been reported from the shale within the quadrangle, but fossils have been found in other parts of this southwest Ozark area. From conodonts collected from a 62-foot section at Spavinaw dam, 10 miles south of the quadrangle boundary, W. H. Hass (written commun., 1952) has identified Upper Devonian species in the lower 46 feet and Lower Mississippian (Kinderhook) species in the top 9 feet; a zone of 7½ feet from which no fossils were obtained intervenes. The Upper Devonian species are common to the Chattanooga Shale of central Tennessee, whereas the Mississippian species are those found in the Maury Formation (Kinderhook) that overlies the Chattanooga unconformably in central Tennessee. Equivalence in stratigraphic position and the lithologic identity of the black shale and its basal phosphatic sandstone over wide areas early led Ulrich (in Adams and Ulrich, 1905; 1911), Moore (1928), and others to correlate the southwest Ozark occurrences with the type area of the Chattanooga, even when the faunas were less well known and more characterized by the similarity in their meagerness than by any positive correlation criterion.

Stratigraphic Relations

The Chattanooga Shale overlies the Cotter Dolomite with major unconformity. Various formations ranging from Early Ordovician to Middle Devonian in age, present farther south in Oklahoma or in northern Arkansas, either were never deposited in the Wyandotte quadrangle or were eroded before the beginning of the Chattanooga Shale transgression. The erosion surface on which the shale was deposited was a level peneplane of very wide extent.

The shale is overlain disconformably by the St. Joe Limestone Member of the Boone Formation. There is no evidence of erosion, unless the sandy shale near the top of the Honey Creek section of the Chattanooga possibly represents a later reworking of the top of the shale.

Mississippian System

Osage and Meramec Series

Boone Formation

The term "Boone" was first published in 1891 in reports of the Arkansas Geological Survey (Simonds, 1891, p. XIII, 27–37; Penrose, 1891, p. 129–138) to designate a heterogeneous unit of cherts and limestones of Mississippian age that crop out widely in northern Arkansas. A basal part, the St. Joe Limestone Member, was differentiated shortly thereafter (Hopkins, 1893, p. 253). The Boone Formation was mapped in the Joplin district folio (Smith and Siebenthal, 1907), and two additional members were described—the Short Creek Oolite Member in the upper part of the formation, and the Grand Falls Chert Member, lying 100 feet below the Short Creek. The part of the Boone between the Grand Falls and St. Joe Members was named the Reeds Spring Limestone Member by Moore (1928). A little later, Cline (1934) proposed that the name Boone, which he considered a synonym of "Osage," be abandoned; that the St. Joe and Reeds Spring Members be raised to the rank of formations; and that the names Burlington Formation, Keokuk Formation, and Warsaw Formation of the standard Mississippian section in southeastern Iowa and adjacent Illinois be extended to the southern Ozark region to apply to equivalent strata included in the Boone Formation. This plan was adopted by Moore, Fowler, and Lyden (1939), though they pointed out that the Burlington was absent in the Tri-State mining district. These authors fitted the informal letter designations of different lithologic units earlier worked out by Fowler and Lyden (1932) into the new classification.

Although the Boone can be divided into several lithologic and faunal units for mapping, this has not been done in the mapping to date (1964) within the Tri-State region, except for delineation of the Grand Falls Chert Member in the Joplin folio (Smith and Siebenthal, 1907), of the Short Creek Oolite Member on the map of Ottawa County (Reed and others 1955, pl. 1), and of the St. Joe unit on the Wyandotte quadrangle map prepared by Siebenthal and Mesler and used in this report (p. 10). Extension of standard Mississippi Valley formation names into the area is not believed to be practical because such extension would have to be on the basis of faunal correlations, and there is as yet no agreement as to where the boundaries should be placed.

The Boone Formation is here retained and is divided into seven members, three of which are new (fig. 2). Although some or all of these could possibly be of formational rank, they have not all been mapped in the Wyandotte quadrangle. Until this is done, all are best retained at the lower rank. The members, their classification in the Mississippian provincial series, and their correlation with the classification of Fowler and Lyden (1932) are given in table 2. Overall thickness of the Boone is about 350–400 feet in the Picher mining field.

TABLE 2.—*Members of Boone Formation in Wyandotte quadrangle correlated with informal letter classification of lithologic units in Picher mining field*

Series	Members of Boone Formation (this report)	Informal letter classification (Fowler and Lyden, 1932; Fowler, 1942)
		Bed
Meramec (Upper Mississippian)	Moccasin Bend Member	B C D E F G H
	Baxter Springs Member	J K L
	Short Creek Oolite Member	M
	Joplin Member	
Osage (Lower Mississippian)	Grand Falls Chert Member	N O P Q
	Reeds Spring Member	R
	St. Joe Limestone Member	

ST. JOE LIMESTONE MEMBER

The St. Joe Limestone Member was formerly exposed at several localities along the Neosho River below the mouth of Honey Creek southwest of Grove and in two tributary hollows (pl. 2). These localities are now flooded by Lake O' the Cherokees. The exposures where the Horse Creek anticline crosses the Neosho north of the Delaware County line have been partly submerged, but at least the upper part of the limestone, and locally all of it, still crops out at the foot of the bluff rising above the lake level and in the adjacent hollows. The St. Joe is exposed where the Horse Creek anticline crosses Buffalo Creek and also at several localities along the Elk River above the Missouri State line.

THICKNESS

The limestone is 32 feet thick on the Neosho River just below the mouth of Honey Creek; it is 14–15 feet thick on the river north of the Delaware-Ottawa County line, and about 21 feet thick in the deep well at the Ballard mine, near Baxter Springs, Kans. It is apparently 10–15 feet thick in prospect drill holes just north of the Kansas line near Picher; but in another drill hole just west of Cardin, typical St. Joe is not present, though what is interpreted as a basal sandstone unit is here 5 feet thick.

CHARACTER

The St. Joe is commonly described as a coarse-grained crinoidal limestone. The matrix, however, is fine grained, and the amount of crinoidal material imbedded in its is variable. Compared to other parts of the Boone, the crinoid stem plates, accounting for the granularity, are rather small. The color usually ranges from light gray to nearly white to drab, and in places the rock has a greenish cast; but the crinoidal crystals may be flesh colored or pink, and if abundant enough, may color the rock pink on fresh exposures. The limestone is laminated in bed 2–18 inches thick. In a soft zone, 3–6 feet thick, whose base is 3–10 feet above the base of the member in the outcrop area, the limestone is earthy, in part nodular, and weathers back in a niche of considerable lateral persistence. Commonly, the softest part of this zone is a few inches of greenish-gray to drab limy shale containing only scattered fine crinoid stem fragments. In general, the soft zone material has fewer crinoidal plates than the rest of the formation. Its bedding may also be irregular or lenticular. Above the soft zone and to the top of the member, the limestone is especially hard and crops out usually in a wall, 7–20 feet high, locally called the "wall limestone." Blebs of amorphous marcasite or pyrite or clusters of fine marcasite crystals are locally present throughout the St. Joe, through they are more abundant in some layers than in others. In some samples this material partly replaces crinoidal fragments.

The St. Joe differs from overlying parts of the Boone in being comparatively free of chert, but locally the basal part, including the soft zone, contains nodules and lenses as much as 4 inches thick, of blue to dark-gray chert; more rarely, the wall-forming unit at the top contains short lenses of chert as much as 6 inches thick.

In the Picher mining field, the basal part of the Boone Formation is penetrated in only a few deep drill holes, commonly those that are drilled into the Ordovician rocks for water. At the Ballard mine near Baxter Springs, Kans., the deep well cuttings assignable to the St. Joe Member are similar in lithology to the rock exposed on the outcrops farther south, except that greenish fine-grained shaly limestone is scattered more or less throughout the 21-foot interval assigned to the St. Joe, instead of being segregated near the base. This shaly material is pyritic in the basal few feet. Four prospect holes that were examined on the Barr tract in Kansas north of Picher cut the St. Joe, one extending through it and barely into the Cotter. The limestone

there is similar to that on the Ballard tract at Baxter Springs. However, three of the holes show partial or complete dolomitization to greenish-gray or pale-brown fine-grained dolomite. As the overlying Reeds Spring is partly dolomitized in some holes but not in others, this dolomite is believed to by hydrothermal and of no stratigraphic significance. Similar dolomite was found in the 5 feet of the St. Joe Member cut at the bottom of a prospect hole on the Gordon No. 3 tract. A little glauconite occurs at the top of the St. Joe in this Gordon hole and in two of the Barr holes. It may be in the dolomite, or in light- to pale-gray chert that occurs sparingly at this level. Glauconite in chert from one of the Barr holes occurs as pale-bluish-green grains, unlike the bright-green glauconite that is common at higher levels in the Boone.

At the base of the Boone where it rests directly on the Cotter in the mining field, as revealed by the deep drill holes, is 3–5 feet of fine-grained sandstone which may be green, pyritic, and variably limy and shaly, grading locally to green shale. The shale may be a thin remnant of Northview Shale, which, according to Lee (1940, p. 30, pls. 6, 7), thins southward in southeastern Kansas but is not present in the Ballard deep well.

Correlation and Age

The St. Joe Member can be recognized by its lithologic character and stratigraphic position over wide areas in northeast Oklahoma, southwest Missouri, and northern Arkansas. Correlation between these different areas is confirmed by the contained fauna. Both faunally and lithologically, the St. Joe resembles the Fern Glen on the east flank of the Ozarks south of St. Louis, and both are classed as the basal units, in their respective areas, of the Osage Series (Moore, 1928). The St. Joe is believed by Moore, Fowler, and Lyden (1939, p. 7) to be the equivalent of the Pierson and Sedalia Limestones in parts of southwestern and central Missouri. Kaiser (1950, p. 2151) has shown, however, that the St. Joe is younger than the Sedalia.

Moore, Fowler, and Lyden (1939, p. 5–6) have classed the more shaly part of the soft zone in the lower part of the St. Joe, as herein described from the outcrop area, as the Northview Shale, and the underlying basal strata as Compton Limestone. These are Kinderhook formations, typically much thicker in Missouri east and northeast of the Wyandotte quadrangle. On the basis of the difference between the St. Joe and Northview faunas and of regional stratigraphic relations, they believe that a disconformity is present between the St. Joe and the underlying Northview.

On the faunal content of the St. Joe Limestone Member, Gordon reports (written commun., 1965), as follows:

> Fossils in six small collections from the lower part of the St. Joe Limestone Member in the Wyandotte quadrangle indicate an early Osage age for this part of the formation. The fossils include *Leptaena analoga* (Phillips), *Chonetes logani* (Norwood and Pratten), *C. multicosta* (Winchell)?, *Productina sampsoni* (Weller), *Marginatia fernglenensis* (Weller), *Rhipidomella jerscycnsis* Weller, *Stenocisma bisinuatum* (Rowley), *Rynchopora persinuata* (Winchell), *Strophopleura novamexicana* (Miller), *Brachythyris suborbicularis* (Hall), and *Cleiothyridina prouti* (Swallow). It is apparent from this list that, in the Wyandotte quadrangle at least, the lower part of the St. Joe Limestone Member is stratigraphically equivalent to the Fern Glen Limestone of Missouri and equivalent beds within the Burlington Limestone, and should not be correlated with the Compton Limestone of Kinderhook age.
>
> Fossils in two small collections from the upper cliff-forming part of the St. Joe, the so-called wall limestone, are similar to those from the lower part of the member. They include *Leptaena analoga* (Phillips), *Chonetes multicosta* (Winchell)?, *Productina sampsoni* (Weller), *Rynchopora persinuata* (Winchell), *Strophopleura novamexicana* (Miller), and *Brachythyris suborbicularis* (Hall).
>
> About 30 species have been identified in the eight collections from the St. Joe Limestone Member in the Wyandotte quadrangle, including corals, bryozoans, crinoids, brachiopods, gastropods, and fish remains. All but the brachiopods are rare.

Stratigraphic Relations

In the few places where the contact has been seen, the St. Joe Member appears to be conformable on the Chattanooga Shale; however, the relation is believed to be one of disconformity (Moore, R. C., and others, 1939, p. 5). Drilling records show that in the underground section within the mining field, the St. Joe rests in places directly on the Cotter. The contact in this area represents a pronounced unconformity, marked by a thin basal sandstone which is the initial deposit of the Mississippian sea that transgressed over the old land surface. This sandstone corresponds to the one at the base of the Boone in other areas on the southwest and south flanks of the Ozarks, which in most places has been called the Sylamore sandstone (McKnight, 1935, p. 67; Moore, R. C., 1928, p. 110).

The St. Joe is overlain conformably by the Reeds Spring Member of the Boone.

REEDS SPRING MEMBER

The Reeds Spring Member of the Boone was formerly exposed southwest and west of Grove in bluffs along the Neosho River below Honey Creek and in adjacent hollows that enter the river from the left, and also in lower Duck Creek and in a middle stretch of Horse Creek,

both of which enter from the right. Most of these exposures are now covered by Lake O' the Cherokees, but possibly some exposures are left in Woodward Hollow or in the river bluff just north of there. Although the top of the Reeds Spring is above lake level in several other places, the upper part of the unit is concealed by slumpage of surface debris from above, except on the steepest slopes subject to most vigorous erosion.

The Reeds Spring crops out for 2 or 3 miles on the Horse Creek anticline in the bluffs and adjacent hollows along the left bank of the Neosho River, beginning 1 mile north of the Ottawa-Delaware County line. Over much of this area the complete thickness of the unit is above lake level. Former exposures along both sides of the river for several miles above this locality to the mouth of Sycamore Creek are now submerged. The member is also exposed where the Horse Creek anticline crosses Buffalo Creek and also rather widely along the Elk River and its tributaries east of the Missouri State line.

Thickness

Because the overlying strata are soft, incompetent, and tend to shed surficial debris down over the Reeds Spring throughout the area of its outcrop, there have been few satisfactory determinations of thickness on the outcrop. Furthermore, the top is gradational into, and interfingers with, the overlying unit, so that there is always a question as to whether equivalent strata are included at the top. At Tynon Bluffs near the southwest corner of the quadrangle a thickness of at least 82 feet assigned to Reeds Spring was formerly exposed above the river level, and there may be other beds below this. In the river bluff 2 miles north of the Delaware County line, the thickness is 70 feet. In the bluff on the right side of Buffalo Creek, about a quarter of a mile above the mouth of Stogdon Hollow, the thickness is at least 96 feet. Deep drilling in the Picher mining field shows a thickness of about 94 feet in the Ballard deep well near Baxter Springs, 105 feet on the Central tract at Cardin, 100 feet on the Gordon No. 3 tract west of Picher, and about 100 feet on the Barr tract in Kansas just north of Picher.

Character

The Reeds Spring Member is typically a thin-bedded alternation of dark chert and fine-grained dark limestone. The proportions vary at different levels, but the chert usually forms 50–70 percent of the total unit. Individual beds range in thickness from 2 to 18 inches and average perhaps 4 to 12 inches. The chert beds commonly have blunt rounded nodular endings against the limestone, and may give way to isolated nodules in certain beds. Paler transitional beds that form the top 14 feet of the member in the bluff of the Neosho, 2½ miles north of the Delaware County line, have irregularly interlocking boundaries between the chert and limestone, with each containing irregular inclusions of the other. Some sections may contain dark-gray shale partings from a fraction of an inch to 2 inches thick.

The chert is characteristically opaque, has a waxy luster, and is grayish blue, gray, or brown or some intermediate hue. Various shades and hues of grayish blue are particularly diagnostic of the Reeds Spring. The brown colors are present in nearly every section, particularly at the top of the member; but they are not so abundant, nor are they distinctive for this unit. The colors may be very dark, but they also commonly grade locally to very pale shades or even white, which in places may be mottled in the darker colors, or conversely, may be mottled by "worm borings" of the darker colors. In this paler material a faint tinge of blue is nearly specific for the Reeds Spring, particularly if it is accompanied by the waxy luster; but the other pale colors are indistinguishable from those in the overlying Grand Falls Chert Member and have the same peculiarities of mottling. The pale cherts are more common in the upper part of the Reeds Spring, but they may occur in the lower part. Although these pale cherts extend roughly along the bedding, they show no stratigraphic persistence that can be recognized over more than very local areas; and adjacent drill holes in the mining field commonly show great differences in the same stratigraphic interval.

In a few places the pale cherts grade, over limited areas, to cotton rock, which is a white siliceous rock intermediate between limestone and chert. Generally, the cotton rock is in thin bands near the top of the Reeds Spring, but it may occur lower in the section. Drill holes on the Barr tract show a thick cotton-rock zone in the upper part of the unit, and some occurs in scattered thin zones to the base. Cotton rock dominates the upper 35–50 feet in some of these holes, forming nearly all of a 12-foot interval in one hole, as well as being conspicuous in an additional 10 feet. For the most part, however, the cotton rock is accompanied by light-gray to nearly white chert which in other places generally exceeds it in volume. The occurrence of such light-colored material in the upper beds of the Reeds Spring on parts of the Barr tract precludes recognition of the top of this unit here, except by comparison with adjacent holes of normal lithology, for the cotton rock is far more abundant in, and characteristic of, the overlying Grand Falls Member.

The limestone of the Reeds Spring is fine grained to dense and gray to brown. Typical colors are rather dark,

but these grade commonly to light and pale shades. In some sections a few thin seams contain coarser grained light-gray to light-brown crinoidal material. Such material appears to be common for a few feet at the top of the Reeds Spring in the southeastern part of the mining field—from the New Chicago, Pat, and Walker tracts to the Betsy Greenback tract north of Quapaw. However, it attains maximum development in the basal Reeds Spring in the southern part of the Wyandotte quadrangle where it appears locally in huge crinoidal reefs, or bioherms (Cumings, 1932). In some of these reefs the limestone is pink, resembling that in scattered beds of the St. Joe Limestone Member.

In a few areas in the mining field the limestone of the Reeds Spring has been altered to a fine-grained dolomite which is usually brown but in places gray. Tracts where this has been noted include the Barr, Pelican, and T. R. Smith (SW¼SW¼ sec. 16, T. 29 N., R. 22 E., Ottawa County, Okla.), and a general area in secs. 10, 11, and 15, T. 35 S., R. 22 E., Cherokee County, Kans. The last two areas are along both sides of the State line several miles west of Picher, and respectively, southeast and southwest of Melrose, Kans. On the Barr tract the dolomite is scattered through the lower half of the Reeds Spring and may alternate with limestone at different levels or may give way to limestone in an adjacent drill hole only 40 feet away. It is in part coextensive with the dolomite in the St. Joe Member, from which it is indistinguishable. On the other tracts the dolomite is limited to the top 15 feet of the Reeds Spring, but on the two tracts in the Melrose area it is associated with a similar or slightly coarser grained dolomite in the Grand Falls. Although this dolomite is unlike the hydrothermal dolomite characteristic of ore deposits at higher stratigraphic horizons, its irregularity of occurrence, lack of stratigraphic continuity, and coextension with similar dolomite in the St. Joe and Grand Falls suggest that it is nevertheless hydrothermal. In one of the drill holes southwest of Melrose, the dolomite is associated with zinc ore just below the top of the Reeds Spring.

Bioherms

Several bioherms in the Reeds Spring Member that were formerly visible along the stream bluffs are now submerged beneath Lake O' the Cherokees, but there are exposures on the Horse Creek anticline where it is crossed by the Neosho River and by Buffalo Creek. The zone of occurrence is at the base of the Reeds Spring, just above the wall-forming unit at the top of the St. Joe Member. The general shape of the bioherms is that of a lens, flat on the bottom and convex above.

A bioherm on the north side of Woodward Hollow about 1 mile above its mouth showed a thickness of 30 feet over an area which was at least 1,100 feet across in the line of the section. This bioherm is made up of coarse crinoidal stem fragments, which, as seen on the weathered surface, are obviously laminated in beds 2–8 inches thick. A little chert in roundish nodules, 2–10 inches across, lies about midway between the top and bottom, but is nowhere as abundant as in normal Reeds Spring. This chert is gray to light gray, light brown, light blue, or has a reddish cast, in many ways resembling the chert in the St. Joe Member more than that in typical Reeds Spring.

The top contact of the bioherm with normal fine-grained and cherty Reeds Spring is fairly sharp. However, thin lenses of the crinoidal material, 1–2 inches thick and commonly only 1–2 feet long, appear in the overlying Reeds Spring, rarely as much as 10 feet above the contact.

On its west flank, the bioherm decreases to a thickness of about 5 feet within a distance of 900 feet, the base remaining flat on the St. Joe contact in this distance and the top dropping accordingly. A short way farther west it has thinned to nothing. Part of the thinning toward the edge is by encroachment of the enveloping normal Reeds Spring through interfingering with crinoidal beds on the flank of the bioherm. In places, isolated lenses of the coarse material as much as 1 foot thick and 10 feet or more long are entirely enclosed in the fine-grained Reeds Spring in the zone of interfingering. An added factor that may contribute to the thinning is the dip of bedding from the center of the bioherm toward the periphery. This dip was not demonstrable in the Woodward Hollow example but could formerly be seen in others.

Peripheral dip may show in all the crinoidal beds on the flank of a bioherm, or only in the upper outer layers that envelop with marked unconformity a central mass of more nearly horizontally bedded or irregularly crossbedded strata. Such features are better illustrated in the bioherms that are thicker and at the same time narrower across the base than the Woodward Hollow example. A splendid example exposed in cross section in the bluff on the left side of the Neosho River, 2 miles north of the Delaware County line, has a thickness of 55–60 feet in the center and a width at the base of about 700 feet (fig. 4). On its northeast flank the crinoidal beds dip fairly regularly toward the periphery from the center of the mass, although there is some overlapping downdip of inner strata by more steeply dipping higher outer strata. The enveloping beds of normal fine-

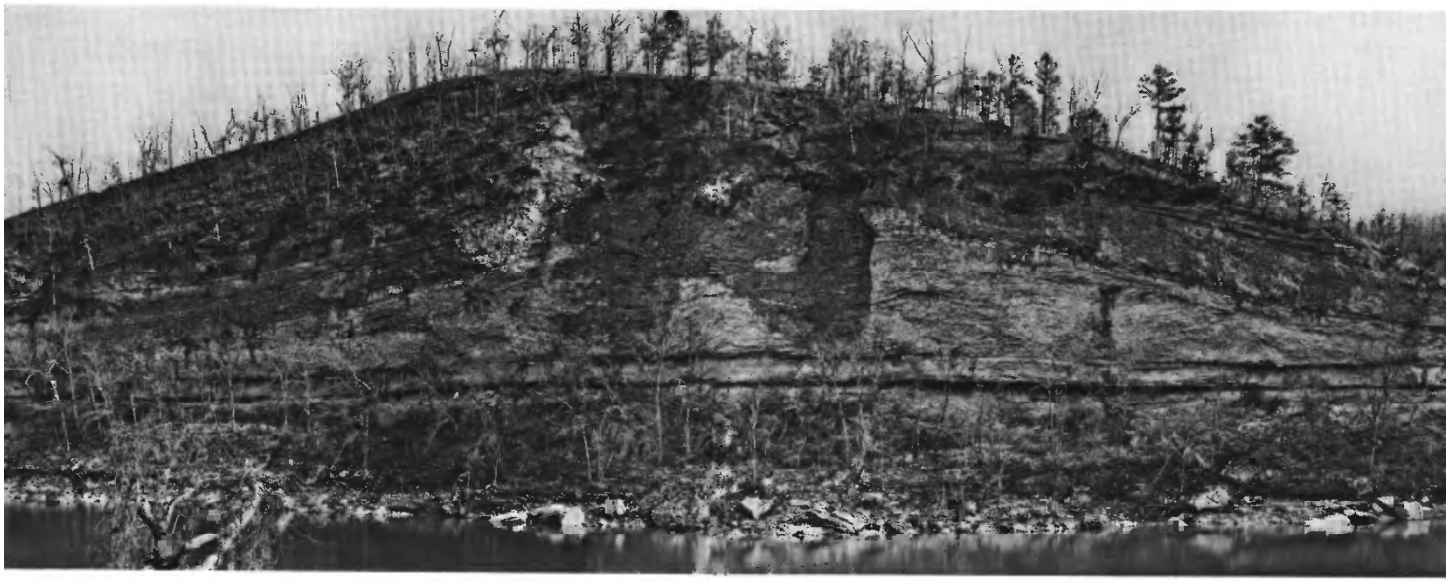

FIGURE 4.—Bioherm, southeast side of Neosho River 2 miles north of Ottawa-Delaware County line; view looking southeast. The massive horizontally bedded unit at base of cliff is the wall-forming bed, here 8 feet thick, that composes the upper part of the St. Joe Limestone Member of the Boone Formation. Photograph by C. E. Siebenthal before impoundment of Lake O' the Cherokees. At present, the top of the St. Joe is a few feet above low-water stage of the lake at an altitude of 735 feet.

grained Reeds Spring limestone and dark chert are conformable but not quite so steeply dipping; the lowest beds lap part way up the flank, and higher beds overlap the lower ones updip toward the summit of the bioherm. On the opposite flank of this bioherm the core consists of small imbricated lenses that, on the average, maintain a fairly horizontal attitude up to the narrow zone where they are abruptly truncated by a thin outer shell of the core material that dips down the slope of the bioherm. The fine-grained typical Reeds Spring strata lie just above and are conformable with the dip of this flank.

This bioherm crops out in a nearly vertical cliff that is for the most part inaccessible. It apparently differs from most others in that the material making up the mass contains a preponderance of chert, which is light colored.[2] The crinoidal composition is largely inferred from analogy with more accessible bioherms, but is partly confirmed by presently accessible exposures near the southwest periphery. Here, in the mouth of the hollow that adjoins the bioherm, the first 7 feet of strata above the St. Joe Member contain scattered fragments of crinoid stems and numerous bryozoan remains in a matrix of limestone and chert that is otherwise typical of the normal Reeds Spring facies.

Perhaps the thickest bioherm in the quadrangle was, before flooding by Lake O' the Cherokees, indicated by incomplete exposures on the left bank of the Neosho River where it crosses the township line near the south boundary of the quadrangle. There, crinoidal material made up 90 percent or more of the limestone in a 74-foot zone just above the St. Joe Member. The strata in this zone were evidently the flank beds of a bioherm, for they showed pronounced fanlike divergence in dip as the still thicker but eroded biohermal core a short distance away was approached; the upper beds dip away from the core at 13° and the lower beds dip away at lower angles. Before erosion of the core, this bioherm must have been nearly 100 feet thick. With normal Reeds Spring strata arching up on its flank, the total Reeds Spring over this bioherm before erosion to the prelake topographic surface probably was considerably thicker than the figures previously quoted, which apply to sections showing only the normal Reeds Spring lithology.

Many other exposures could formerly be seen in which coarse crinoidal beds containing variable amounts of nodular or bedded chert (2–3 in. beds) alternate with fine-grained limestone and chert in the transition zone

[2] Part of the description given herein is based on the notes of Siebenthal and Mesler. These notes, in describing the different lithologies, are rather vague and not entirely consistent, probably because of inaccessibility of the outcrops.

at the base of the Reeds Spring. Such exposures were undoubtedly in the fringe areas of bioherms.

Exposures on the bioherms have not been complete enough to reveal their true three-dimensional shapes. Possibly the long flat section of the Woodward Hollow example may represent a prevailingly longitudinal section of a mass that is noticeably elongate in shape, whereas the other exposures showing the steeper dips on the flanks are more nearly cross sections at right angles to the elongation. However, the possibility is not ruled out that the prevailing shape may be nearly circular in plan, as most sections show a relatively narrow apex instead of a long flat one. Figure 5 shows diagramatically the relations of a bioherm to the enclosing strata.

In lithology, the bioherms resemble the St. Joe Member much more closely than the Reeds Spring Member, and they thus might be considered as being an upward extension of St. Joe in certain reeflike bodies into the Reeds Spring. Indeed, they are classed as St. Joe by Laudon (1939, p. 326–327, 338). However, the stratigraphic continuity of the wall-forming unit, which elsewhere marks the top of the St. Joe, beneath the bioherms shows that the bioherms are later than the rest of the St. Joe and are equivalent in age to the basal Reeds Spring elsewhere, into which they interfinger. They undoubtedly grew as submarine mounds on the floor of the early Reeds Spring sea, for the fine-grained limestones and dark cherts that normally were deposited in this sea show only partial overlap on the flanks of the mounds, and are draped over their crests in anticlinal arches. The broken crinoidal material composing the bioherms appears to be entirely detrital and stratified, though commonly very irregularly, in all the exposures examined. That it was not evenly spread over the floor of the early Reeds Spring sea by contemporary current action was presumably due to the weakness of such currents in deep water. There was some stratification of the crinoidal material in comparatively steep slopes on the flanks of the mounds, and some spreading of the material as thin layers in the lime muds adjacent to the mounds, but apparently the material was not spread any great distance from the mounds.

In the Paleozoic seas, as in modern ones, wave-resistant reefs composed of the skeletal remains of many types of organisms were able to grow from bases below the influence of wave action into the surface zone of destructive wave action through the ecologic adaptation of a relatively few groups of organisms (Lowenstam, 1950). Such reef-building organisms were not only able to withstand the conditions of the surf, but they acted as sediment binders on the unconsolidated debris from all sources on which they grew. Other groups of organisms were passive occupants of the reef biotic community, contributing nothing to the structural rigidity of the reef, but contributing much skeletal debris as a building material for utilization by the active reef builders. Surrounding the rigid cores of the mature reefs were steeply dipping flank strata, forming a talus fan of bioclastic debris that was derived from comminution, in the wave zone, of the calcareous skeletons of all reef inhabiting types. The framework of the reef builders is subordinate to the debris from other types of reef organisms, and is not commonly exposed in outcrops of fossil reefs (Ladd, 1950, p. 204). During the Paleozoic the reef builders were certain groups of colonial corals, the stromatoporoids, and probably algae, whereas the chief suppliers of debris were the crinoids (Lowenstam, 1950, p. 435–449; Goldring, 1938, p. 51–67).

In the bioherms of the Wyandotte quadrangle the clastic nature of the crinoidal debris and the irregularity of bedding that is common in this coarse-grained bioclastic material indicate derivation by vigorous wave action near the surface of the sea. The limited size of the bioherms, their isolation, and the peripheral inter-

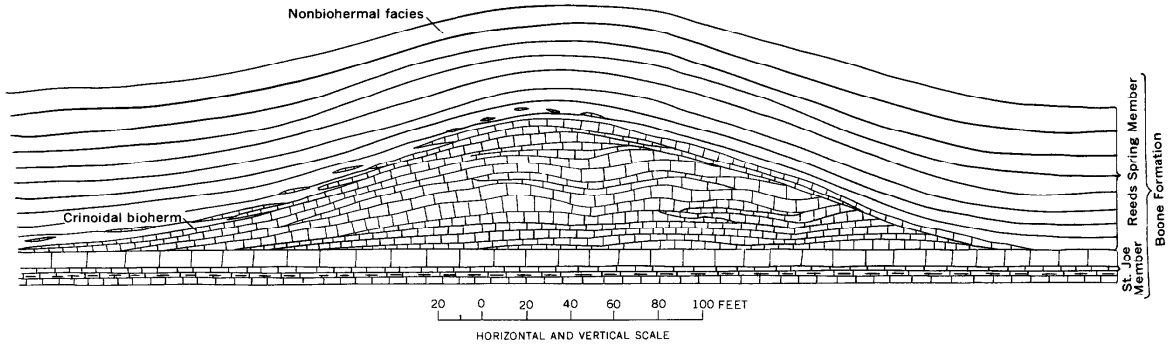

FIGURE 5.—Sketch of cross section showing relations of a bioherm to enclosing strata in basal part of Reeds Spring Member of Boone Formation, based on observations at several localities.

fingering of their flank beds with the horizontally bedded fine-grained facies of Reeds Spring limestone which surrounds and covers them preclude explaining them as being due to a general shoaling of the contemporary seas to a depth less than effective wave base. On the other hand, the characteristics of the exposed bioherms are compatible with an interpretation of them as phases of isolated reefs whose foundations were below effective wave base but whose crowns projected well above wave base. Some of the bioherm exposures appear to represent steeply dipping detrital flank beds, others could be tangential cross sections of flank beds farther removed from the wave resistant core. The bioherm now showing in the bluff above Lake O' the Cherokees, 2 miles north of the Delaware County line (fig. 4), may represent a cross section closer to a reef core. No critical study has been made of the bioherms to determine whether remains of reef builders, which would characterize the reef cores, are recognizable.

Growth of the reefs was presumably terminated by deepening of the Reeds Spring sea at a rate faster than the reef-building organisms could raise their base. The darkness that comes in this latitude at perhaps 300–400 feet of depth in the sea would have exterminated the reef-building organisms and arrested any further growth of the reefs (cf. Lowenstam, 1950, p. 477).

Correlation and Age

The type locality of the Reeds Spring is in Stone County, Mo., 60 miles east of the Wyandotte quadrangle. Correlation with the type locality is based on similarity in lithology and stratigraphic succession in the two areas. Numerous exposures in the intervening region are known (Moore, R. C., 1928, p. 190), and the outcrops in the two areas could be connected by mapping with, at most, very small gaps.

Fossils are not common in the Reeds Spring Member, but they have been found, chiefly in the lower part, in both the normal fine-grained limestone and in the coarser grained lenses constituting the bioherms in the southern part of the quadrangle. The fauna is dominated by brachiopods, bryozoans, and crinoids; though the last are not represented by identified species, the crinoidal bioherms are evidence of their importance in the life of that time. Gordon comments (written commun., 1965) as follows:

> Fossils in 13 collections from the Reeds Spring Member in the Wyandotte quadrangle total a little over 50 species, including corals, bryozoans, blastoids, crinoids, brachiopods, pelecypods, gastropods, and trilobites. Brachiopods by far dominate the fauna, but crinoid remains are locally abundant in bioherms at the base of the member. Six of the collections are from these bioherms in Woodward Hollow, a tributary of the Neosho River now inundated by Lake O' the Cherokees. All the collections came from the lower 30 feet of the member.
>
> Among the 30 species of brachiopods in the fauna are **Rhipidomella oweni* (Hall and Clarke), **Schizophoria swallovi* (Hall), †*Chonetes arkansanus* Girty, var., †*C. multicosta* (Winchell), *Quadratia* sp., †*Geniculifera* cf. *G. boonensis* (Branson), *Rhytiophora* n. sp., **Echinoconchus vittatus* (Hall), *Setigerites newtonensis* Moore, *Rhynchopora persinutata* (Winchell), **Spirifer logani* Hall, *Imbrexia buehleri* (Moore), **I. incerta* (Hall), *Brachythyris suborbicularis* (Hall), *Strophopleura novamexicana* (Miller), *Punctospirifer subtextus* (White), **Pseudosyrinx missouriensis* Weller, *Torynifer cooperensis* (Swallow), *Athyris lamellosa* (L'Eveille), and †*Cleiothyridina glenparkensis* Weller. Those marked by an asterisk (*) are restricted to the crinoidal bioherms, and those marked by a dagger (†), to the normal fine-grained facies. The rest are common to the two facies.
>
> The crinoidal bioherms in the lower part of the member, among the masses of crinoid debris, contain, besides brachiopods, a few corals, bryozoans, and many snails of two or three species of the platycerid type. Each platycerid snail lived on the calyx of a crinoid, nested between its arms and perched over the animal's anal orifice (Bowsher, 1955, p. 2–6). Some of these snails exceed 60 mm in length and obviously were associated with large crinoids whose columnals reach 25 mm in diameter, and whose branching lower parts of the stalk that stuck down into the mud reached 40 mm in diameter.
>
> Oddly enough, no crinoid calyxes have been recovered in the biohermal facies. This may be in part because of recrystallization of much of the calcite in the crinoid plates and columnals, together with the susceptibility to weathering of the crinoidal masses, which has obliterated details of the surface in two specimens that appear to be basal parts of crinoid calyxes. But more probably it is because the large crinoids are of a type whose calyxes disintegrated readily after the death of the animal. Several specimens of a small species of *Synbathocrinus* have been found in the fine-grained limestone near the edges of the bioherms.
>
> Most of the fossils listed from the Reeds Spring Member are typical Burlington forms. This was recognized at an early date by Stuart Weller to whom Girty sent one of the Woodward Hollow collections (USGS loc. 1146–PC) for examination. Weller (letter to Girty dated May 25, 1914) pronounced the fossils to be clearly of Burlington age.
>
> The problem that yet remains to plague us is whether or not rocks of late Burlington age are represented in the Reeds Spring. The collections from the Wyandotte quadrangle are not entirely helpful in this respect because they come only from the lower third of the member. A collection reported by R. C. Moore (1928, p. 190–192) to have come from 10 feet below the base of the Grand Falls Member, which overlies the Reeds Spring Member, on Shoal Creek in the Joplin district, Mo., contains many elements in common with the fauna from the lower part of the Reeds Spring Member in the Wyandotte quadrangle. Thus, indications are that the fauna is fairly uniform throughout the member.
>
> R. C. Moore (1928, p. 192–193) regarded another fauna, from crinoidal limestone near Crane, Mo., as representing a higher horizon than the one on Shoal Creek and correlated it with the upper part of the Burlington. Since then, some geologists insist on the presence of a significant unconformity between the Reeds Spring and Grand Falls Members of the Boone Formation. This

idea finds further support in the presence of typical Burlington Limestone above as much as 40 feet of chert and limestone similar to that of the Reeds Spring in sections farther northeast in Missouri, such as the one at Castlewood near St. Louis.

Nevertheless, the absence of any physical field evidence for an unconformity at the top of the Reeds Spring Member indicates the need for another explanation of the facts. It seems entirely possible that the seeming absence of the upper part of the Burlington in the Picher and Joplin mining fields is due to its replacement laterally by a fine-grained limestone facies that has brought with it the brachiopod fauna normally regarded as more typical of the lower part than the upper part of the Burlington. As R. C. Moore (1928, p. 193) has pointed out, the most important among the faunal elements of the Burlington Limestone are the crinoids. In the Wyandotte quadrangle as well as in the Joplin area, no crinoids have been collected from the upper part of the Reeds Spring, though crinoidal material has been found at this horizon in churn-drill cuttings in a small part of the Picher field (p. 23).

True, there is a considerable change in the brachiopod fauna at the base of the overlying Grand Falls Member, but this same change occurs in the type Mississippian section at the base of the rocks now included in the Keokuk Limestone. Moreover, a number of Burlington species carry over into the Grand Falls Member and die out there. Crinoids in U.S. Geological Survey collections from Grand Falls Member equivalents in northwest Arkansas are Burlington types, according to Kirk (oral commun., 1941). For these reasons the St. Joe and Reeds Spring Members in the Tri-State region are regarded as representing Burlington time approximately in its entirety, but because of facies differences, subdivision of these rocks into lower part and upper part of the Burlington as in the type Mississippian section is not possible.

Stratigraphic Relations

No evidence has been seen to indicate that the Reeds Spring Member is other than conformable on the St. Joe Member, although R. C. Moore (Moore, R. C., and others, 1939, p. 8) has indicated a probable unconformity at this horizon.

The relation of the Reeds Spring to the overlying Grand Falls Member is subject to wide differences in interpretation. In our experience, a gradational conformable contact with interfingering in the lithologies of the two members is most consistent with the physical evidence. This contact is discussed more fully on pages 31–32.

GRAND FALLS CHERT MEMBER

The Grand Falls Chert Member formerly cropped out in a dendritic pattern along drainage lines over a wide area in the Wyandotte quadrangle downstream from the junction of the Spring and Neosho Rivers. Much of the best outcrop was drowned by Lake O' the Cherokees, but there are still many places along the upper reaches of the streams tributary to the Neosho from the southeast, along Horse Creek near its junction with Fly Creek, and on upper Hickory Creek, where the unit would be exposed if it were not so incompetent physiographically. Because much of the siliceous material in the unaltered formation is calcareous and easily weathered, the unit acts like a shale through much of its outcrop area, and good exposures are not common. The best exposures were formerly in Tynon Bluffs, and in a bluff on the right side of the Neosho River west of the mouth of Council Hollow, but these are now submerged or partly submerged.

The unit has also been extensively explored in drill holes throughout the mining field. Because the Grand Falls has proved to be ore bearing in widely scattered areas, many of the holes drilled since 1930 were not stopped until the dark-colored Reeds Spring cherts underlying the Grand Falls had been reached.

Thickness

The Grand Falls Chert Member is 88 feet thick in Tynon Bluffs, 75 feet thick in the bluff on the right side of the Neosho west of the mouth of Council Hollow, and 80 feet thick in Missouri on the right side of the Elk River, just below the mouth of Blackfoot Branch.

In the mining field the thickness ranges from 25 to 95 feet. As the base is gradational to, and interfingers with, the underlying Reeds Spring, it is very probable that the thicker sections ascribed to the Grand Falls overlie correspondingly thinner sections of Reeds Spring; but too few drill holes reach the base of the Reeds Spring to demonstrate this conclusively. On the tracts for which a reasonably uniform thickness (94–105 ft) has been previously cited for the Reeds Spring, the Grand Falls shows the following thicknesses (the range in holes that do not reach the base of the Reeds Spring is shown in parentheses): in the Ballard deep well, 60 feet; on the Central tract at Cardin, 52 (50–60) feet; on the Gordon No. 3 tract west of Picher, 40 (32–49) feet; and on the Barr tract north of Picher, 47–54 feet. In some of these Barr holes, the top of the Reeds Spring is established by a thin zone showing typical Reeds Spring chert colors over an interval of 5–12 feet, but below which the Grand Falls type of lithology extends for variable distances. Had the top Reeds Spring layer been missing in all holes, as it is in some, the Grand Falls would have been considered much thicker on the Barr. Thicknesses of 87–95 feet that have been logged on the Harris tract, 2,000–3,500 feet to the southeast, may well correspond to a stratigraphic section in which the upper Reeds Spring strata are indistinguishable in lithology from the Grand Falls.

When the cuttings from a series of holes on any given tract are studied, the thickness of the Grand Falls varies perceptibly, perhaps even between adjacent holes. The variation within a 40-acre tract may be as much as 20 feet, though it is usually less.

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Thickness of the Grand Falls throughout the mining field in not everywhere known, but certain regional variations are indicated that may be significant in interpreting the stratigraphic history of the Osage Series in the general region. For much of the mining field, the thickness is between 45 and 65 feet. Widely scattered tracts on which most thicknesses fall within this range include the Lafalier, Walton, H. W. Smith, and Coffey tracts on both sides of the State line southwest of Melrose, Kans.; the T. R. Smith tract (SW¼SW¼ sec. 16, T. 29 N., R. 22 E.) in Oklahoma southeast of Melrose; the Laura Jenny Zheka and northeast forty of the Mehunka Zheka Beaver tracts just south of the State line, 2–3 miles west of Picher; the Little Greenback (SW¼NE¼, sec. 24), Velie Lion, John Beaver, Crystal-Central, Ritz, and north halves of Blue Goose No. 2 and See Sah tracts in the vicinity of Cardin; the Walker and Maxine tracts, 1 mile southeast of Picher; the Barr tract in Kansas just north of Picher; the Burns (SW¼ sec. 25, T. 34 S., R. 23 E), Thompson (NE¼SE¼ sec. 36, T. 34 S., R. 23 E.), and Garrett mine tracts, 3 miles north of Picher; the Thomas (NW¼NW¼ fractional sec. 15, T. 35 S., R. 24 E.) and Ballard mine tracts, 1–2 miles southwest of Baxter Springs; and probably the Crane tract, 2 miles northeast of Quapaw.

There are two areas in which the thickness averages less than 45 feet. One is in the mineralized ground southwest of Melrose, Kans., in a belt north and northeast of the area of "normal" thicknesses previously indicated, and includes parts of the Porter, Von Treba, Vanatta and A. T. Wright tracts where the thickness is 37–45 feet. The other area is in a zone extending from northwest to southeast across the center of the Picher mining field. Tracts (with thicknesses) within this belt include the Federal Jarrett, 35–49 feet; Stanley, 42–47 feet; Pelican, 30–49 feet; Gordon No. 3, 32–49 feet; Dorothy Bill No. 2, 39–45 feet; Kenoyer (NW¼SW¼ sec. 20), 30 feet; Cardin townsite, 37–40 feet; Barbara J., 30–51 feet; Acme, 42 feet; New Chicago, 35–45 feet; Pat, 40–42 feet; and Betsy Greenback tract (just north of Quapaw), 25–45 feet. Not enough drilling through the Grand Falls has been examined to define the northeast edge of this belt in Treece, Kans., and north Picher. A single hole on the Big Elk tract shows a thickness of 40 feet, and it is quite possible that this may lie within the belt. The southwest edge is better defined by drilling on the Admiralty, which shows thicknesses of 40–50 feet, by the drilling in the vicinity of Cardin, previously cited, which shows normal thicknesses of 45–65 feet, and by the drilling on the See-Sah and Blue Goose tracts, which show transitions to the thick Grand Falls sections at the southwest end of the Picher field.

There are two areas in which the Grand Falls thickness will average in excess of 65 feet. Of three pertinent holes examined on the Scammon Hill tract at the southwest side of the field, one just barely reaches the base at a thickness of 85 feet, and the other two have not reached it at 87 feet. The north edge of this thick zone has been fairly closely defined by drilling on the See Sah and Blue Goose No. 2 tracts, where an irregular line can be drawn near the north side of the south tier of forties, separating thicknesses of 65–76 feet on the south from thicknesses of 51–65 feet on the north.

The other general area of thick Grand Falls is on the north side of the field in Kansas, and includes the Harris tract, 87–95 feet; Earl Coe tract (NE¼NW¼ sec. 4, T. 35 S., R. 24 E.), 65 feet; Swalley tract (NW¼NW¼ sec. 3, T. 35 S., R. 24 E.), 67–75 feet; and Iron Mountain tract, 60–78 feet. There is, however, a hole between the last two mentioned (in NW¼SW¼ sec. 3) in which the thickness is only 52 feet; and it is quite possible that had a larger part of the drilling on the north side of the field been studied, a complicated variation in thickness might have been detected.

Character

The siliceous constituent of the Grand Falls Chert Member varies in different areas or in different parts of a given section from a hard, brittle "butcher knife" flint to a soft, porous cotton rock or tripoli containing a variable but commonly very minor amount of microscopic interstitial calcite. The cotton rock is white and is usually intermixed with opaque pale chert into which it grades imperceptibly, with no discernible boundary pattern. An intermediate product is a soft chert with the texture of unglazed porcelain. In its outcrop area within the Wyandotte quadrangle, the cotton-rock phase predominates in some areas and the pale cherts in others. Examination of drill cuttings shows that the cotton-rock phase with the associated pale chert prevails on the unaltered fringes of the mining field and also in some large relatively barren blocks within the field, such as on the Dorothy Bill No. 2. In most of the mines, however, the hard flinty chert is the characteristic form.

Most sections contain a little limestone which may make up 10–40 percent of certain zones, commonly 15–20 feet thick, though other zones of comparable thickness have none. For the whole unit the limestone probably averages about 10 percent. In the hard chert area within the mining field, the limestone tends to be absent or less common at the top of the Grand Falls than lower

in the section. Other limestone-free zones have no recognizable stratigraphic persistence. Although the limestone may be sharply separated from the cotton rock, there are also complete gradations from the one to the other, and the intermediate products are difficult to identify properly. Locally, the cotton rock or cotton-rock-like chert may show the granular texture of the associated limestone which it has evidently replaced. In parts of the mining field, all the limestone was leached from the Grand Falls, probably during the period of emergence between the Mississippian and Middle Pennsylvanian.

The Grand Falls Member is in beds that are usually 3–18 inches thick, though the limestone is commonly in lenticular beds that may be as thin as 1 inch interstratified with the chert or cotton rock. The chert may appear also as lenses, egg- or kidney-shaped nodules, irregular nearly equidimensional nodules, or fine marbling in certain beds of the limestone. The chert may, in turn, contain irregular inclusions of the limestone with ragged indefinite boundaries.

The harder cherts of the mining field range from opaque to translucent, and are dominantly light to pale gray, grading to nearly white, and light to pale brown, with a little full brown in places. Many sections also have light to pale-blue chert which is usually more translucent than the blue chert of the Reeds Spring, although there are types that are indistinguishable. Medium-gray chert is found in a few places but always as a minor constituent. The paler colors may form a rim, half an inch or more wide, bordering the outer surfaces of the beds and nodules.

The more colored opaque cherts commonly show a very fine speckling of the coloring material in a paler groundmass. This speckling appears to be a relict texture from replacement of finely granular limestone. It appears in places in cotton rock. There is also a coarser mottling ($\frac{1}{16}$–$\frac{1}{8}$ in. diameter) of dark-gray, brown, or dark-brown spots in the paler opaque cherts, or locally in cotton rock, which we have termed "coach dog" mottling. It may be clustered in irregular groups 1–2 inches across, and though it resembles worm borings, it is believed to be relict from the replacement of the stemlike parts of bryozoans. As it is especially characteristic of the massive chert (L bed of Fowler and Lyden) in the Baxter Springs Member at a higher zone in the Boone, it will be described more fully in the treatment of that unit (p. 41). Still another type of mottling, especially characteristic of the flinty translucent chert at the top, takes the form of pale or nearly white opaque rims surrounding an irregularly rounded dark spot 1–2 inches across. As these pale rims are similar to those that commonly surround clusters of "coach dog" mottling, they may have a related origin.

The limestone in the Grand Falls is typically fine grained and brown to gray with intermediate hues, grading to pale colors especially where associated with cotton rock. Any given sample can be matched by the limestone at some place in the Reeds Spring, but on the average, the Grand Falls contains more brown and less gray, and is likely to be lighter colored. In a few places a little medium- to coarse-grained limestone may be included, generally as grains in the finer material, but no stratigraphic significance can be recognized. Some of the medium-grained material shows a granularity that appears to be related to recrystallization. In the mineralized area along the State line southwest of Melrose, Kans., and also on the T. R. Smith tract southeast of Melrose, most of the limestone has been altered to brown, or locally gray, dolomite. Most of this is fine grained, though in places it is medium or even coarse grained.

The top of the Grand Falls in a large part of the mining field is a distinctive type of translucent hard chert, pale-gray or pale-brown, grading in places to light shades of these colors, though appearing darker in massive outcrop because of the translucency. It commonly contains dark spots or "eyes," 1–2 inches in diameter, surrounded by opaque pale rims, but does not have the other types of Grand Falls mottling. In places, this chert may lack its usual translucency. It is well bedded, and typically has no interbedded limestone, although scattered lenses 1–3 inches thick are present. Its lower limit is indicated by the appearance of the other types of chert, even though the type found at the top may persist as one of the components; and there is commonly a roughly concomitant appearance of appreciable limestone or its mineralized equivalent, jasperoid, in the section. The base of the upper phase is indefinite, as the more colored opaque cherts, or lenses of limestone, appear higher in some sections than in others. This is reflected in the range in thickness, from 3 to 30 feet, of the upper chert phase, the average being perhaps 15–20 feet. In some sections, and particularly on the fringes of the mining field where the hard chert gives way to the cotton-rock-bearing phase, the upper part of the Grand Falls is indistinguishable from the rest of the member. It would thus appear that in the mineralized ground, the upper part of the Grand Falls has undergone "chertification," if by this term is understood any process by which nearly 100 percent chert is derived from an original material of something less than 100 percent chert. The change from cotton rock to translucent chert involves little or no chemical mass exchange

and might be conceived as, or comparable to, a late stage in diagenesis. The loss of the interbedded limestone is, however, not so simply explained. The limestone in the top part of the Grand Falls amounts to about 10 percent in much of the fringe area, but in some holes it may reach 30 percent, rarely more.

The only other units of at least pseudostratigraphic appearance recognizable in the Grand Falls are ill-defined and nonpersistent cotton-rock-bearing zones that occur in the prevailing chert over small areas, as delineated by drilling in the mining field. In addition to such local occurrence in recognizable zones, cotton rock may occur haphazardly and generally as a subordinate constituent anywhere in the cherty phases of the Grand Falls, even in the hard flinty chert at the top.

The Grand Falls Chert Member includes the N, O, P, and Q beds of Fowler and Lyden (Fowler, 1942, p. 207). The flinty translucent chert at the top is for all practical purposes their N bed, although they have designated by this letter the stratigraphic interval, 20–30 feet thick, rather than the lithologic unit. The other three subdivisions are also stratigraphic intervals differentiated because they have been mineralized differently in different areas. However, they show no distinctive lithologic features by which they can be distinguished. The Grand Falls Chert Member is the host rock for extensive "sheet ground" ore deposits in areas southwest of Baxter Springs, Kans., and south of Cardin, Okla.

Correlation and Age

The type locality of the Grand Falls Chert Member is a few miles southwest of Joplin, Mo., and about 10 miles east of Baxter Springs, Kans. Because of the general westerly dip, the top of the Grand Falls lies about 200 feet below the surface at Baxter Springs. However, it is easily recognized in drill cuttings, and its top was used as the datum plane for the structure contours shown in the Joplin folio, which includes Baxter Springs in the southwest corner (Smith and Siebenthal, 1907).

The unit herein classed as Grand Falls has been studied in many drill cuttings from the Iron Mountain tract less than 2 miles southwest of Baxter Springs. It shows the same lithologic character and stratigraphic position, including the same interval below the Short Creek Oolite Member, as described in the Joplin folio, and the altitude of its upper surface, 585–635 feet above sea level, is consistent with the 620-foot contour shown in the Joplin folio on top of the Grand Falls at the west side of Baxter Springs. Although the thickness on the Iron Mountain tract (60–78 ft.) is somewhat greater than that ascribed to the Grand Falls around Joplin (35–55 ft.), these latter figures are closely matched in other parts of the Picher field, as previously indicated (p. 27–28). Finally, the extensive development of typical sheet-ground ore in the Grand Falls as herein correlated on the Ballard and adjacent tracts, just northeast of the Iron Mountain tract, is further evidence that the correlation with the sheet-ground formation near Joplin is sound.

On the fauna and regional correlation of the Grand Falls Chert Member, Gordon (written commun., 1965) reports as follows:

> The fauna of the Grand Falls Member includes corals, bryozoans, brachiopods, and rarely mollusks and trilobites, some 60 species in all. These were identified in 20 fossil collections from the Wyandotte quadrangle. This fauna is characterized by the earliest appearance of the following brachiopods, generally regarded as typical Keokuk forms: *Schizophoria compacta* Girty, **Orthotetes keokuk* (Hall), **Avonia williamsiana* Girty, * *Echinoconchus biseriatus* (Hall), * *Labriproductus wortheni* (Hall), * *Marginatia? crawfordsvillensis* (Weller), *Rotaia subtrigona* (Meek and Worthen), * *Rhynchopora palmeri* Girty, * *Spirifer rostellatus* Hall, *S. tenuicostatus* Hall, * *Pseudosyrinx keokuk* Weller, * *Torynifer pseudolineatus* (Hall), * *Eumetria verneuiliana* (Hall), and *Cleiothyridina parvirostris* (Meek and Worthen). Those species marked with an asterisk (*) range upward into beds of Warsaw age.
>
> One of the most characteristic of these fossils is *Rotaia subtrigona*. This relatively large rhynchonelloid branchiopod ranges through the beds generally referred to the Keokuk Limestone and its equivalent and appears to be restricted to them. Its stratigraphic range in the Wyandotte quadrangle includes the overlying Joplin Member of the Boone Formation.
>
> The Grand Falls Member also contains several Burlington species that are not found in the overlying Joplin Member. These include *Marginatia fernglenensis* Weller, *Stenocisma bisinuatum* (Rowley), *Cyrtina burlingtonensis* Rowley, *Punctospirifer subtextus* (White), *Torynifer cooperensis* (Swallow), *Hustedia circularis* (Miller), *Ptychospira sexplicata* (White and Whitfield) and *Athyris lamellosa* (L'Eveille).
>
> Several other brachiopods, about half of them new, appear to be restricted to this member, including *Tylothyris similis* (Weller), *Cyrtina neogenes* Hall and Clarke, *Composita pentagonia* Weller, and *Beecheria sinuata* (Weller). The little spiriferoid *Tylothyris similis* is exceedingly abundant in the upper part of the Grand Falls Member, including the chert ledges on Shoal Creek that form the falls at Grand Falls, near Joplin, and it appears to be restricted to this part of the section.
>
> The Grand Falls Member is approximately equivalent to the Lower Keokuk of Van Tuyl (1925, p. 47, 146), the "cherty beds of passage" of some authors, also called "Montrose chert" by others. This part of the type section of the Mississippian System was regarded as separate and distinct from both Keokuk and Burlington by some authors and included in the Burlington by others. Van Tuyl pointed out that some Keokuk species appear for the first time in these beds and therefore referred them to the Keokuk, in which he is followed by the U.S. Geological Survey.

Stratigraphic Relations

The Grand Falls overlies the Reeds Spring with an apparently conformable gradational contact or with interfingering between the two. As the limestones are very similar in the two members, their distinction is based on the difference in color of the chert. Typical Reeds Spring chert is dark and characteristically blue, but the color fades irregularly toward the top, less commonly downward, and stratiform intercalations of typical Grand Falls cherts may be included in the Reeds Spring.

Figure 6 shows the distribution of chert colors in selected drill holes that lie in an irregularly sinuate southwest-to-northeast section from near Cardin to near Baxter Springs. The chert layer of Grand Falls type shown near the top of the Reeds Spring in Gordon No. 3 hole F233 is recognizable in most of the drill holes examined on that tract. In hole F253, however, the chert in this interval has the full Reeds Spring colors. Holes F240 and F250 show thinning or paling of the Reeds Spring layer overlying the zone of Grand Falls type, and in hole F241, lying nearly on a line between the two last mentioned, the upper layer of Reeds Spring has so changed in chert color as to be indistinguishable from Grand Falls.

The four Barr holes included in figure 6 illustrate clearly the difficulty in interpreting the boundary between Reeds Spring and Grand Falls. Holes 237 and 232 show thin zones of typical Reeds Spring lithology at approximately the level that marks the top of this unit in many other parts of the mining field, but in each there is an underlying interval of Grand Falls lithology, dominantly cotton rock in hole 237 and chert in hole 232. Below this interval of typical Grand Falls lithology in hole 232, the chert is predominantly of Grand Falls type but shows its affinity to the Reeds Spring in the presence of some pale-blue and light-bluish-brown chert. In the other two holes the zone of Reeds Spring lithology at the top of the Reeds Spring Member is lacking, though there is higher occurrence of sparse Reeds Spring colors in the Grand Falls, recognized in all four Barr holes. As previously pointed out (p. 29), the Grand Falls commonly contains some light to pale-blue chert that is indistinguishable from lighter phases of the Reeds Spring, and it is possible that such material may be in part the stratigraphic equivalent of the uppermost Reeds Spring in other areas where the Grand Falls is very thin. Where the Grand Falls apparently thickens abruptly, the thickening undoubtedly

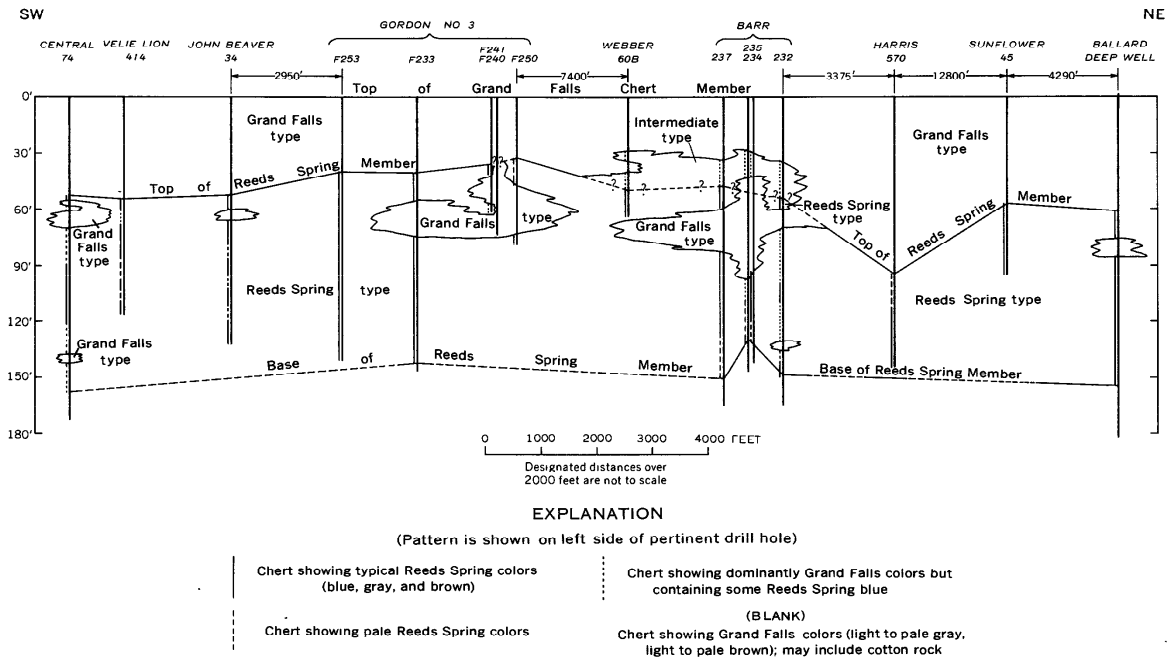


Figure 6.—Diagram showing distribution of chert colors in Grand Falls Chert Member and Reeds Spring Member of Boone Formation in selected drill holes between Cardin and Baxter Springs (SW–NE). Note alternation and intergradation of Grand Falls and Reeds Spring types. Details of the intertonguing shown are conjectural.

has been effected by color fading of Reeds Spring that elsewhere extends higher in the section. Thus, fading of the Reeds Spring "islands" in dominantly Grand Falls lithology in the two holes on the Barr (237, 232) would produce the situation illustrated in Harris hole 570 (fig. 6), which is very similar to that in the other two holes on the Barr (234, 235). Unfortunately, the Harris hole did not reach the base of the Boone; therefore the presumed thinning of the Reeds Spring cannot be demonstrated.

If the common alternation and intergradation of Reeds Spring and Grand Falls lithologies in vertical section is not kept in mind, plotting of the contact as determined from drill cuttings would suggest an extremely irregular contact characteristic of erosional unconformity. However, the true relation is believed to be one of conformity. The interpretation herein deduced from a study of well cuttings in the mining field is borne out by examination of the contact in the few places where it is, or was formerly, exposed in the southern part of the quadrangle.

The Grand Falls at its type locality is considered by Cline (1934, p. 1142) to be "only a local chert variant of the upper Reeds Spring"; and Laudon (1939, p. 328), in ascribing a thickness of 186 feet to the Reeds Spring "along Grand River southwest of Grove," has obviously included in the upper part of his Reeds Spring unit the strata that we have classed as Grand Falls in the Lake O' the Cherokees impoundment basin.

The Grand Falls is overlain with apparent conformity by the Joplin Member.

JOPLIN MEMBER

The Joplin Member, which is a chert-bearing limestone, crops out widely in the Wyandotte quadrangle—along the Neosho River and its tributaries from near Mudeater Bend to the southwest corner of the quadrangle and up the Spring River to the Kansas State line. The top of the member was never very high above the streambed of Spring River, and even dipped below river level on parts of Moccasin Bend, and possibly elsewhere. Lake O' the Cherokees has further restricted outcrops along the two rivers above their junction, but the top of the member is above streambed at the head of the lake on Spring River, and is some 50 feet above the lake level on the point between Spring River and Shawnee Branch. East of Spring River the strata rise on approximately the same dip as the gradients of the tributary streams, so that the Joplin Member is exposed along several miles of Fivemile Creek to the vicinity of Hornet, and along Lost Creek at least to the spring a mile east of Racine, nearly at the east boundary of the quadrangle.

Southward from the junction of the Neosho and Spring Rivers, the strata rise relative to the drainage lines onto the flanks and crest of the Horse Creek anticline, and the Joplin Member is widely exposed along both the Neosho and its tributaries from the east, to and including the Elk River and its tributaries. Downstream from the mouth of the Elk, the unit dips below the level of the lake and even below the former bed of the Neosho River; but it rises again west of Grove and crops out extensively, its base lying near lake level along lower Honey Creek and rising gradually eastward up this creek. The unit also crops out in the hills bordering parts of Duck and Horse Creeks on the west side of the Neosho, particularly where the latter creek is crossed by the Horse Creek anticline.

As the Joplin Member has been the major ore-producing stratum in the Picher mining field, it has been extensively cut in the underground workings and in the many prospect holes that have been drilled in that general area. It corresponds in effect to the M bed of Fowler and Lyden (Fowler, 1942, p. 207 [3]). Although their M bed also includes the overlying relatively thin Short Creek Oolite Member, this oolite is so universally leached, squeezed, condensed, and replaced in the mineralization process that it becomes indistinguishable at the top of their M bed unit, if indeed, it has not been completely removed in most mineralized sections.

THICKNESS

The Joplin Member varies greatly in thickness, owing to several independent factors. There were undoubtedly initial differences in the total accumulation of strata in different areas, probably with a diastemic break at the top of the unit. There may also have been slight erosion locally at the top of the unit, possibly by submarine truncation in shallow water. As the Joplin Member is the most plastic and structurally incompetent unit in the section, there has undoubtedly been much local thickening and thinning in response to diastrophic stresses in certain areas. And last, because the limestone of the Joplin Member has been extremely soluble in underground solutions, including the mineralizing solutions, the thickness is in consequence subject to wide and locally abrupt variations due to solution and slumpage.

Exposures examined on the surface can be selected to avoid the latter two factors, which are secondary postdepositional processes, so that only the first two, which may be classed as primary stratigraphic factors, are

[3] In the several papers by Fowler and Lyden and by Fowler, the stratigraphic concepts differed somewhat over the years as more and more was learned about the stratigraphy of the district. Citation is, therefore, to one of the later papers.

applicable. In one of the best exposures found, in a short steep gorge locally known as Wildcat Hollow near the top of the bluff on the right side of the Neosho River, west of Council Hollow (probably in fractional section 20, T. 26 N., R. 24 E.), the thickness is 76 feet. In the bluff on the left side of the Neosho River, 2 miles southwest of Wyandotte, a thickness of 48 feet was formerly exposed, which included all the Joplin Member except perhaps 2 or 3 feet of the marker bed at the base; a thickness of 50 feet can be accepted for this locality. On the south side of the Elk River, 2 miles above its mouth, the member as formerly exposed was at least 70 feet thick, and on the west side of Spring Branch near its mouth, northwest of Grove, it was at least 56 feet thick. Southwestward from here the thickness decreases irregularly and locally, owing to the disconformity at the base of the next overlying unit. Thus, at former outcrops, now below lake level, in a north-facing bluff 3 miles west-northwest of Grove, the Joplin Member is about 25 feet thick; but within a distance of 600 feet the unit is beveled by the disconformity to a thickness of only 10 feet at the structurally high point, beyond which the Joplin wedge increases again in thickness, though irregularly. In Tynon Bluffs, near the southwest corner of the Wyandotte quadrangle, the thickness is about 10 feet. On Honey Creek below the mouth of Elm Creek, 2 miles southwest of Grove, the thickness has increased to 49 feet. Most other exposures in the southern part of the quadrangle fail to show the full thickness of the unit, usually because the upper part has weathered back into a concealed slope.

A study of stratigraphic thicknesses in the mining field involves the examination of cuttings from many churn-drill holes. In individual holes, the secondary variations in thickness that are superposed on the original stratigraphic variations are not distinguishable in drill cuttings if due to structural deformation, and only partly distinguishable if due to solution and slumpage. Errors can usually be avoided, however, by considering several adjacent holes. Where the rocks are unaltered in the fringe areas of the mining field or in some of the barren areas between deposits, maximum thickneses of the Joplin Member are not greatly different from the Wildcat Hollow or Wyandotte sections. Considerably thinner sections that from study of drill cuttings appear to be equally unaltered may represent an initially thinner unit, but also may locally record condensation by stylolitic solution or more extensive leaching along bedding seams. Residual greenish clay is very common in the limestone sections, but as it is usually washed out of the drill cuttings, no evidence is left of this particular clue on which to judge how much condensation has taken place. However, an increase in the percentage of insoluble chert with decrease in the thickness of the member can be demonstrated in many instances and is believed to be adequate proof that at least many of the variations below the maximum, particularly within short distances, are due for the most part to condensation by solution. Hence, only the maximum thicknesses observed in local areas would appear to have any stratigraphic significance.

The stratigraphic thickness of the Joplin Member in the Picher mining field ranges from 0 to 100 feet. In a line of properties that roughly encircles the field from the southwest side around the west and across the north to the northeast side, maximum thicknesses observed in drill cuttings examined by us, which may approximate original thicknesses of the Joplin Member, are as follows; Scammon Hill, 56 feet; See Sah (southeast forty), 70 feet; Blue Goose No. 2 (south forty), 72 feet; Wilson, 57 feet; Little Greenback (southwest forty), 70 feet; Laura Jenny Zheka, 82 feet; Federal Jarrett, 55 feet; Semple, 56 feet; Early Bird (NW¼NW¼ sec. 12, T. 35 S., R. 23 E., Kans.), 70 feet; Garrett (NW¼ sec. 36, T. 34 S, R. 23 E., Kans.), 72 feet; Barr, 77 feet; Harris, 67 feet; Dobson, 63 feet; Swalley (NW¼NW¼ sec. 3, T. 35 S., R. 24 E., Kans.), 80 feet; Iron Mountain, 100 feet; and Ballard, 70 feet.

Less of the drilling was studied on the south and southeast sides of the field, and possibly the fringe areas comparable to those cited for the other sides were not reached. Maximum thicknesses in what appear to be unaltered sections, judging from the prevalence of unaltered limestone, are as follows: Crane, 43 feet; Betsy Greenback, 35 feet. There has been no demonstrable concentration of chert in the limestone of these two tracts; indeed, on the Greenback tract, the percentage of chert in many holes is less than in other areas of considerably greater unit thickness. Hence, it is probable that the somewhat smaller thicknesses quoted for these two tracts may represent the true original thickness. Although structural thinning is not ruled out, the approximate uniformity in thickness over a considerable area makes it unlikely.

A similar situation at a few places nearer to the center of the mining field suggests that the Joplin Member originally may have been thinner in part of the field than on the fringes. Thus, along the north side of the Lucky Syndicate east forty, several holes in apparently unaltered limestone show a thickness of only 22–34 feet; and over a large part of the John Beaver east forty, an equally unaltered section shows thicknesses of 45–18 feet, decreasing from south to north.

Just north of this John Beaver tract, in the southeast forty of the Anna Beaver, there is a particularly puzzling area in which underground examination shows that the Joplin Member is absent with no apparent local complication that would explain it on structural grounds. Contouring of thickness on the two tracts demonstrates a progressive thinning northward to the area, which is perhaps 500 feet across (fig. 7), showing complete cutout on the Anna Beaver. This truncation is undoubtedly related to the disconformity at the base of the overlying unit.

The abrupt thinning of the Joplin Member from unaltered to mineralized ground, due to leaching and collapse, can be demonstrated at innumerable places. A typical example is between holes 147A and 145A on the Laura Jenny Zheka, where an unaltered limestone section, 72 feet thick, decreases within 250 feet to an altered (jasperoidized) section only 30 feet thick. The base remains at a nearly constant altitude in this distance, and the top drops by approximately the amount of the thinning. Similar relations are shown between Blue Goose No. 2 holes A74 and A64, where the thinning, again by sagging of the upper surface, is from 60 to 30 feet within a distance of 280 feet; at an additional distance of 250 feet is the edge of a large stope. Between holes 98C and 97C on the Scammon Hill tract, the Joplin Member similarly thins from 56 to 25 feet within a distance of 100 feet. Abrupt thinning may also occur without alteration of the limestone, for example, in Blue Goose No. 2 hole A38 where the unit is 31 feet thick and is only 230 feet from hole A87 in which the unit is 66 feet thick. In some sections that have been thinned, part of the limestone appears to be recrystallized. On the other hand, the unit may be altered and mineralized in many places, particularly on the fringes of the mining field, without marked previous condensation. Thick mineralized intervals in the Joplin Member are common in such widely separated mines as the Wilson (57 ft), Pelican (70 ft), Barr (55 ft above an additional basal 20 ft), and Iron Mountain (80 ft).

In the heart of the mining field where comparatively little unaltered limestone remains, probably all the factors that affect the thickness of the Joplin Member have been effective, either at different places or at the same place at different times. Although involvement of the different factors can usually be recognized, their relative importance can rarely be appraised. The resultant thickness of the Joplin Member varies widely, but from 10 to 25 feet is a common range. Thicknesses of less than 10 feet commonly result from the concentration of structural gliding within the incompetent Joplin Member, with attendant brecciation and squeezing, followed by later solution. In some structurally deformed areas, as on the axis of the Rialto trough along the east line of the Rialto tract, the unit may thus be reduced to zero thickness.

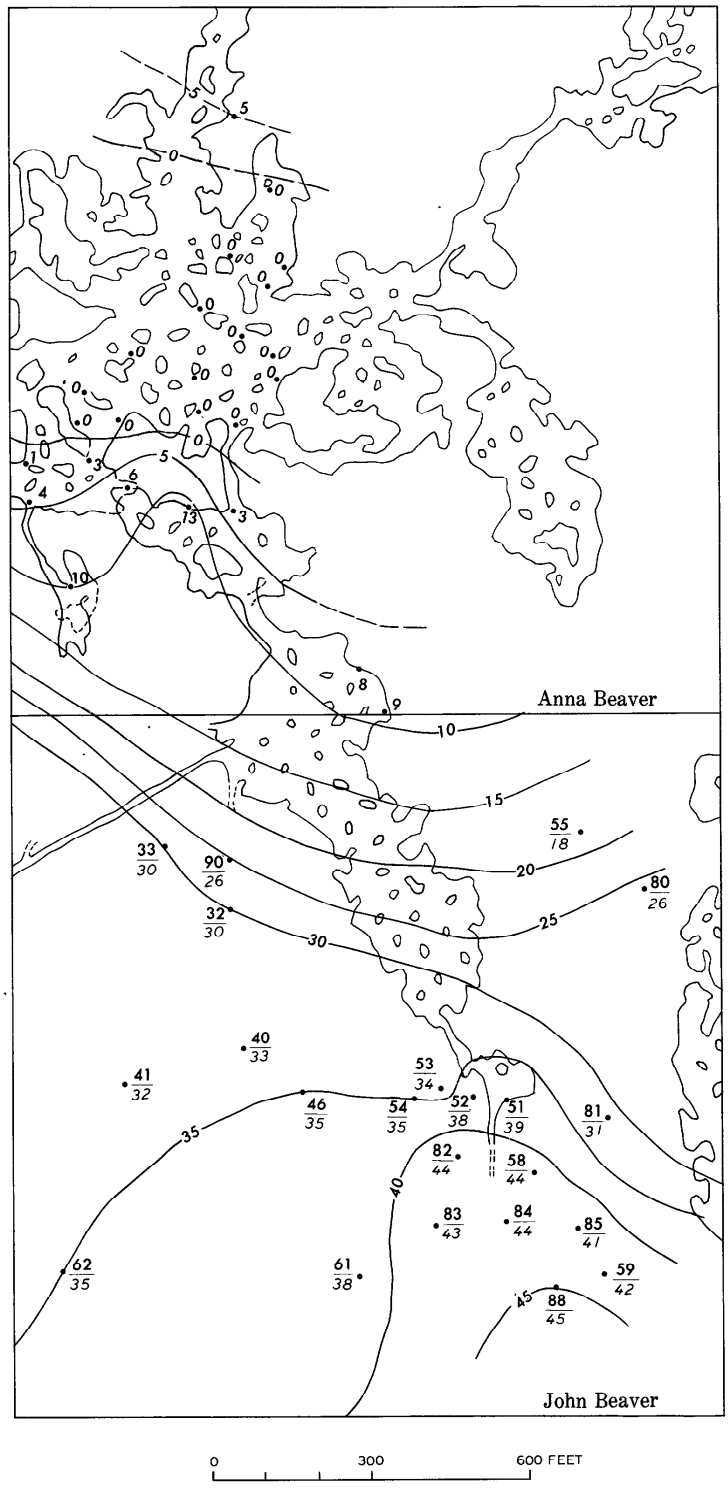
In the mineralized area along the State line southwest of Melrose, Kans., the Joplin Member ranges in thickness from 40 feet in unaltered sections cut by drill holes on the H. W. Smith tract (E½W½ fractional sec. 15, T. 35 S., R. 22 E., Cherokee County, Kans.), to 0 on C. A. Coffey and Porter land, a quarter of a mile to the northeast, and also to 0 on at least some of the Vanatta land 1 mile east. The truncation of the Joplin Member is undoubtedly stratigraphic, owing to the disconformity at the base of the Short Creek, but within the wedge of truncation, the thickness shows the usual additional variation due to secondary solution effects. The area of Coffey and Porter land on which there is no Joplin Member is apparently not more than about 1,800 feet across, as a short distance to the northeast (Von Treba land) the Joplin has increased to apparently unmodified thicknesses of 30 and 35 feet. It is probable that the areas on which there is no Joplin on the Coffey and Vanatta tracts are connected in a strip north of the State line.

Between the Melrose area and the Picher field, an apparently unmodified original section of the Joplin Member is 23 feet thick in one of the holes examined on the Whitby tract (W½W½ fractional sec. 16, T. 35 S., R. 23 E., Cherokee County, Kans.).

Character

The Joplin Member in unaltered sections is a coarse- to rather fine-grained crinoidal fossiliferous limestone, thick-bedded to massively bedded, and except in a zone near or at the base, containing appreciable chert and locally a little cotton rock. The chert may be in beds from 3 inches to 3 feet thick, in long lenticular nodules, or in watermelon shaped or rounded nodules from a few inches to perhaps 2 feet thick. It may also occur in less regular masses with indefinite boundaries, usually elongated along the bedding directions. Such chert commonly contains indefinite blebs or larger irregular residuals of the limestone. The cotton rock, which is present only in association with unaltered limestone, occurs in masses of the same size and shape as the chert, into which it usually grades, or it may be intimately intermottled with the limestone. In places it preserves the granular texture of the replaced limestone.

The limestone is gray to light gray, in part with brownish or bluish tinge; less commonly it is brown or light brown. It may be sparingly glauconitic in scattered strata, and is commonly stylolitic. In part of the miner-



EXPLANATION

25

Isopach showing thickness of Joplin Member below horizon of Short Creek Oolite Member of Boone Formation
Dashed where inferred. Interval 5 feet

9

Thickness of Joplin Member, in feet
On John Beaver tract, member is unaltered and below unaltered oolite. In Anna Beaver mine, member is altered and mineralized, and oolite is destroyed

88

Drill-hole number

Mine workings
Dashed line indicates different level of mine

N

FIGURE 7.—Isopach map of Joplin Member of Boone Formation below horizon of Short Creek Oolite Member on parts of John Beaver and Anna Beaver tracts.

alized area southwest of Melrose, Kans., the limestone has been greatly thinned and altered to fine-grained brown dolomite.

The chert is typically opaque, gray to dominantly light gray, also light brown or light blue, grading commonly to pale shades or nearly to white. There may be darker mottling on a coarse irregular scale, or on a finer scale roughly comparable to the granularity of the associated limestone. Some of the mottling is of coach dog type (see p. 41), usually coarser and more diffusely scattered than in the Baxter Springs Member. Some of the bedded cherts have scattered dark irregular "eyes" surrounded by pale rims, similar to those more characteristic of the upper chert in the Grand Falls Member. The chert nodules may be banded concentrically in pale and darker shades, with a thick nearly white gradational band at the periphery. The beds of chert likewise show pale borders 1–2 inches thick. In these borders, which may grade to cotton rock, the chert commonly contains microscopic interstitial calcite grains, and various types of stylolites may have developed locally in such limy chert. Rarely, the center of a chert nodule, or some of the concentric bands, may be unreplaced limestone. The chert commonly contains silicified fossils of various types—bryozoans, brachiopols, corals, crinoid stems, or microfossils—and there are also many unreplaced calcite crinoid stems.

The brecciation of the chert in the mining field and elsewhere is a structural feature related to deformation of the limestone and is discussed more fully on page 77–78.

A characteristic and widespread unit near or at the base of the Joplin Member is chert free or nearly so. This unit crops out widely in the southern part of the Wyandotte quadrangle as a rounded ledge, commonly 10–20 feet thick, and makes an excellent marker bed. It is also recognizable in uncondensed sections penetrated in drill holes on the fringes of the mining field, where it is as much as 30 feet thick, though commonly only 15–20 feet. In both the outcrop and mining field areas, this unit may form the base of the member, but is is as commonly underlain by 5 feet to as much as 15 feet of chert-bearing limestone at the base of the member. Considering the irregularity of chert occurrence in the Joplin Member, it is probable that the base of the member is everywhere contemporaneous and that initial deposition of the limestone was accompanied by chert in some places but not in others. Even the "chert free" zone commonly contains a little chert, and locally, it contains enough to lose its distinctive character.

Above the nearly chert-free unit the Joplin Member contains a variable amount of chert, irregularly distributed, to the top of the unit. Although the percentage may be small in some sections at certain levels, nearly every 5-foot interval contains some chert, separating the limestone into beds from 3 inches to 4 feet thick, averaging perhaps 1–2 feet. The amount and distribution of chert vary considerably from place to place, but in most of the apparently unleached sections the chert in the interval above the basal unit will average 15–50 percent, having this range in different drill holes on the same mining tract.

Any leaching, to which the Joplin Member is especially susceptible, concentrates the chert to a higher percentage of the unit, usually with a contraction in thickness. The chert-free zone at the base is particularly vulnerable to leaching in the mining field and is destroyed in most thinned sections. In the extreme situation, all or practically all the limestone is removed, and even the chert may show such evidences of solution as irregularly and poorly formed stylolites. The resultant texture in such condensed material is similar to the headcheese texture (described on p. 42) in the lower unit (L bed) of the Baxter Springs Member. In the average mineralized section in the heart of the mining field, chert forms 70–90 percent of the Joplin Member. The chert-free zone at the base may be represented by a residual clay layer; or the zone may have been the focus for cave development in the erosion interval that preceded the Pennsylvanian, with the later introduction of muds that formed the Pennsylvanian shales at this level when the land was again submerged.

Correlation and Age

The Joplin Member is herein named for its occurrence in the quarry of the Joplin Marble Quarries Co., on the left bank of Shoal Creek half a mile or so below Grand Falls, 3 miles southwest of Joplin, Mo. (Hinchey, 1946, p. 38). Here at the type locality, the "chert free" marker bed near the base of the member is the principal stone that is quarried. According to observations made by Mackenzie Gordon, Jr., the marker bed is here 19 feet thick, though split by a thin cotton-rock parting, and its base lies 18 feet above the base of the member. The total Joplin Member, which is about 78 feet thick at the quarry site, has the same stratigraphic sequence of lithologies and the same relative position between the underlying Grand Falls Chert Member and overlying Short Creek Oolite Member as shown by drilling in the Picher mining field, a few miles to the west.

The Joplin Member contains abundant fossils, though the number of species represented is not commensurate with individual abundances. Fragments of crinoid stems are common, but species are rarely identifiable. On the

faunal content, Gordon reports (written commun., 1965) as follows:

> The fauna of the Joplin Member, as shown by a study of 17 fossil collections from the Wyandotte quadrangle, includes about 40 species of corals, bryozoans, brachiopods, and rare mollusks and tribolites. This fauna numerically is smaller than that of the Grand Falls Chert Member because of the absence from it of the Burlington and restricted species that occur in the Grand Falls. The entire fauna of the Joplin consists of species that range upward from the underlying member. *Setigerites setiger* (Hall) is fairly common in the upper part. This species is very rare in the Grand Falls Member and has been found in it only in the Joplin district. *Schizophoria compacta* Girty, *Rotaia subtrigona* (Meek and Worthen), and *Spirifer logani* Hall do not occur above the top of this member.
>
> As none of the species in the Joplin Member is restricted to it, it is the strong Keokuk flavor of the fauna and the absence in it of any characteristic Burlington or Warsaw elements that typify the Joplin fauna. The Joplin Member thus correlates with the type Keokuk section at Keokuk, Iowa, which is the upper part of the Keokuk Limestone of Van Tuyl (1925, p. 142–154).

Stratigraphic Relations

The Joplin Member overlies the Grand Falls Chert Member with apparent conformity, although the marked change in the lithology must record an abrupt change in the conditions of deposition. It is overlain disconformably by the Short Creek Oolite Member of the Boone.

SHORT CREEK OOLITE MEMBER

The Short Creek Oolite is a thin but persistent bed of oolitic limestone that crops out widely along the Neosho River and its tributaries, formerly from the first bend above Mudeater Bend to Honey Creek, and along Spring River and its east tributaries as far north as the Kansas State line, and east to Hornet and Racine. Its outcrop is virtually coextensive with that of the Joplin Member north of the Horse Creek anticline. Many former exposures near or at stream level on Spring River and on the Neosho above its junction with Spring River have been submerged by Lake O' the Cherokees, but the oolite is above lake level at least near the mouth of Shawnee Branch. Several good high exposures are along the Neosho from the mouth of Spring River to just below the mouth of the Elk River, traversing the flanks and crest of the Horse Creek anticline. Much of the outcrop belt in southeastern Ottawa County is shown on plate 1 of Oklahoma Geological Survey Bulletin 72 (Reed and others, 1955), but that along Spring River and its tributaries in the northeastern part of the county is not indicated.

South of the Horse Creek anticline, along the Neosho River and lower part of the Elk River, the Short Creek Oolite Member was not always recognizable where the top of the Joplin Member was exposed. It is not evident whether this is due to a failure in primary deposition at such localities, or whether the oolite has here been destroyed by recrystallization. No occurrences have been noted in the Wyandotte quadrangle southwest of the stretch of river extending from the mouth of Honey Creek northwest to the bend east of Needmore, except for a 1-foot bed in the bluff east of Needmore which has a vague suggestion of oolitic texture. Former exposures of oolite along the river between the latter locality and the mouth of the Elk are now below lake level, but in places, typical occurrences could be seen to grade within short distances into crystalline limestone in which no trace of oolite could be recognized. Good exposures remain above lake level along Honey and lower Elm Creeks southwest of Grove and along many parts of the Elk River and its tributaries eastward well into Missouri.

The Short Creek Oolite Member is present throughout the mining field, as far west as the mineralized area along both sides of the Kansas-Oklahoma line at the northwest corner of the Wyandotte quadrangle, and north to the Garrett mine, 3 miles north of Picher. It shows in most drill holes that penetrate unaltered limestone sections; but as the oolite is very susceptible to recrystallization and leaching, it is not generally recognizable in mineralized ground, except where it contains sparse oolitic chert preserved in the ore breccia. The oolite forms the top of the M bed of Fowler and Lyden (Fowler, 1942, p. 207).

Thickness

The Short Creek ranges in thickness from 1 to 10 feet, but in most sections it is from 4 to 6 feet thick. At its southwesternmost occurrences in the quadrangle, it is 2 feet thick on the north side of Honey Creek half a mile below the mouth of Elm Creek; the apparently equivalent interval in the bluff east of Needmore is only 1 foot thick. This thinning at the most southwesterly occurences known in the quadrangle suggests that the edge of the oolite has been reached in this direction, and that its absence farther southwest is due to nondeposition rather than to some form of destructive alteration. Eastward from this fringe zone the unit reaches a thickness of 4 feet on Elm Creek within 1 mile of the fringe locality cited, of 3–3½ feet opposite the mouth of Hickory Creek, 4 miles northeast of the Needmore locality, and of 5 feet in the river bend a couple of miles farther northeast. The increase to the northeast up the river is not regular, however, for the oolite is only 1½ feet thick at one place between the last two localities; and as previously pointed out, there are gaps in which the oolite is missing, including local-

ities even farther northeast along the lower course of the Elk River. In scattered localities within the area of normal thickness the oolite is less than 4 feet thick, as on Spring River at the mouth of Shawnee Branch where it is only 2 feet thick.

An area of maximum thickness lies near the junction of the Neosho and Spring Rivers. The oolite is 9 feet thick in an old quarry three-quarters of a mile southwest of Wyandotte; and 6–10 feet thick at different places on the rock spur between the two rivers. In these thicker sections the oolitic texture may be poorly developed in part of the interval, either at the top or bottom of the unit. In the bluff of the Neosho, 1½ miles southwest of the mouth of Sycamore Creek, the thickness is 7 feet; on the point between Spring River and the mouth of Warren Branch, the thickness is also 7 feet; and at several localities near Seneca, the thickness is 8 feet.

Outcrops where the thickness falls within the 4- to 6-foot interval are too numerous to mention, and the cuttings from drill holes within the mining district suggest that this thickness is also about average for that area. As the drill cuttings in unmineralized ground are bailed out every 5 feet according to standard drilling practice in the field, it is not possible to estimate thin units such as the Short Creek very closely. For the Dobson and Federal Jarrett tracts, thickness estimates as much as 10 feet appear to be good, and are maximum for the mining field. On many tracts the Short Creek appears to be less than 4 feet thick; but as there is some possibility of condensation by solution, the minimum is not very definite.

Character

The Short Creek is a very even textured oolite. The spherules are thickly packed and mostly rounded, averaging about one-fiftieth of an inch in diameter, but a few are oval, oblong, or less regular. Some of the oolite is slightly glauconitic, especially toward the top, and the glauconite grains tend to lie in the centers of oolite spherules. The color of the unit is brown to light brown, in contrast with the usually gray or brownish-gray limestone of the Joplin Member that is below it. On the weathered surface the oolite commonly is pale brown or pale gray, approaching a chalky white that makes it conspicuous at a distance. It is typically massive, though some bedding may be visible in less oolitic material at the top and bottom. Locally, as in the left bluff of the Neosho River 2 miles below the mouth of the Elk River, the unit may be crossbedded at low angles. It crops out in a slope or niche, commonly concave, and characteristically scales off in flakes parallel to the surface of exposure.

In many sections small indefinite blebs of the oolite have been altered to gray, brown, or light-blue to nearly white chert, which preserves the light-brown silicified spherules usually within an opaque whitish matrix. In rare occurrences the oolitic chert is nodular, as at one locality just above lake level on the Spring River side of the point between the Spring and Neosho Rivers, where such a nodule, 34 inches long and 19 inches wide and well banded near its periphery, is oriented perpendicular to the plane of stratification (fig. 8).

The Short Creek Oolite Member is very unstable and readily loses its distinctive texture by recrystallization or solution. The recrystallized unit becomes fine-grained limestone that commonly retains its brown or light-brown color after all traces of oolitic texture have been destroyed, thus furnishing a clue as to its identity. As previously noted, the oolitic texture comes and goes in the thinned unit as its southwesternmost occurrences are approached. Possibly, this may be an original depositional feature, but it could represent recrystallization of the type that is so common elsewhere. The oolite was very soluble in the mineralizing solutions, and it is

FIGURE 8—Banded nodule composed of oolitic chert, in Short Creek Oolite Member of the Boone Formation, on point between Spring and Neosho Rivers.

doubtful that much of it remains as a stratigraphic unit in the mines, though its horizon was commonly the locus of rich mineral deposition. Where the oolitic limestone has been removed, any small amount of oolitic chert that was present in the Short Creek is concentrated as a residue at or near the top of the M bed. It is more likely to be found as sparse fragments in the drill cuttings of holes that happen to intersect such chert remnants, but a careful search at the proper horizon underground will commonly reveal it, if present.

Correlation and Age

The type locality (Smith and Siebenthal, 1907) is on Short Creek at Galena, Kans. about 7 miles northeast of Baxter Springs. As the unit is lithologically unique and areally persistent, it is a marker bed to which other parts of the Boone are commonly tied stratigraphically. The occurrence in cuttings from churn-drill holes on the Ballard and Iron Mountain mining tracts within 2 miles southwest of Baxter Springs shows the same relation to underlying strata and approximately the same interval above the Grand Falls Chert Member as in the type area. This interval decreases somewhat to the west in the mining field and southward in the outcrop area in the south half of the Wyandotte quadrangle.

Gordon comments as follows on the contained fauna (written commun., 1965):

> Study of 13 collections from the Short Creek Oolite Member shows that its fauna consists of about 40 species of fossils, including rare corals, fairly common bryozoans, dominant brachiopods, and a few mollusks, trilobites, and fish remains. Among the bryozoans *Cyclopora fungia* Prout is one of the most common forms. The brachiopods include a number of late Osage species that range upward from the beds below, such as *Orthotetes keokuk* (Hall), *Labriproductus wortheni* (Hall), *Echinoconchus biseriatus* (Hall), *Rhynchopora palmeri* Girty, *Spirifer tenuicostatus* Hall, *Torynifer pseudolineatus* (Hall), and *Eumetria verneuiliana* (Hall). With these species are associated early Meramec species, including *Perditocardinia* n. sp., *Marginirugus magnus* (Meek and Worthen), *Tetracamera arctirostrata* (Swallow), *Composita globosa* Weller, and *C. trinuclea* (Hall). The Warsaw pelecypod *Pseudaviculopecten amplus* (Meek and Worthen) also occurs in the oolite.
>
> *Tetracamera arctirostrata*, described originally from the oolitic Salem Limestone, appears to be restricted in the Wyandotte quadrangle to the Short Creek Oolite Member, where it is common. *Marginirugus magnus*, on the other hand, has not been found in true oolite lithology but is restricted to light-gray medium-grained limestone associated with the oolite. Along Spring River, at least from Moccasin Bend northward to the Kansas State line, a 1-foot bed of this limestone, carrying *Marginirugus magnus*, occurs between the coarser grained medium-gray limestone of the Joplin Member below and oolite above. As the oolite grades laterally into similar rock locally, the *Marginirugus*-bearing limestone is believed to belong in the Short Creek Member. This is the earliest known occurrence of *Marginirugus magnus* in the Wyandotte quadrangle and is believed to mark the base of the Warsaw in this region. The Short Creek is thus the basal rock unit of the Meramec Series.

Stratigraphic Relations

The Short Creek overlies the Joplin Member with apparent disconformity, to judge by the large variation in the stratigraphic thickness of the Joplin Member in different areas. In the Joplin region the interval between the Grand Falls Chert Member and the Short Creek is 100 feet (Smith and Siebenthal, 1907), and this interval is approximated on the Iron Mountain tract near the northeast edge of the Picher mining field. To the southwest in the mining field, the Joplin Member occupying this interval is truncated beneath the Short Creek Oolite Member and is locally cut out on a part of the Anna Beaver tract, 5 miles from the Iron Mountain. Although the stratigraphic relations have been obscured by the leaching and condensation to which the limestone units have been subjected in the mineralized areas, a study of drill cuttings from the less altered blocks of ground in which there is no apparent concentration of chert reveals numerous intermediate sections. The ground in and near the known area of complete cutout has been mineralized, and hence the oolite has here been destroyed; but the progressive thinning of the Joplin Member beneath oolite in unaltered ground approaching the area of complete cutout is convincing evidence that the base of the Short Creek is the horizon of truncation (fig. 7).

A truncated wedge of the Joplin Member in the Melrose area, decreasing in thickness from 40 feet to 0 within a quarter of a mile, has been discussed in the section on thickness of the Joplin Member. Undoubted Short Creek has not been recognized on top of the Grand Falls Chert Member beyond the edge of the Joplin wedge in this area, but as all limestone at its horizon has been leached or altered to jasperoid, its absence is explainable. The Short Creek is present in one hole near the point of the wedge above only 5–7 feet of unaltered Joplin-type limestone that contains a normal 30 percent of chert.

The belt of thinned or truncated Joplin beneath the Short Creek in both the Picher field and near Melrose coincides roughly with the areas of thinnest Grand Falls below the Joplin, previously pointed out (p. 28). This fact suggests the interesting possibility that there may have been a structural or submarine topographic swell in the floor of the Mississippian sea, that traversed the Picher-to-Melrose axis and lasted through an appreciable period of Osage time. If so, the sedimentary hiatus between the Grand Falls and Joplin Members might, at least along this axis, represent more of a break than elsewhere.

Another area in which the Joplin Member is truncated beneath the Short Creek, previously pointed out (p. 33), is along the Neosho River west-northwest of Grove. The oolite is present at one place where the Joplin is 25 feet thick but is not recognizable 600 feet away where a bed at its horizon has beveled the underlying Joplin to within 10 feet of its base. Nearly 2 miles farther west in the bluff east of Needmore, a thin and slightly glauconitic bed with a vague suggestion of oolitic texture may represent an altered remnant of the Short Creek, which here ovelies at least 17 feet, and probably more, of the Joplin. All these outcrops are now submerged. Farther southwest at Tynon Bluffs, only the "chert-free" marker bed, about 10 feet thick at the base of the Joplin Member, remains, but the oolite is not recognizable at the base of the overlying truncating strata.

The truncation at the base of the Short Creek resembles an erosion unconformity, yet the top of the underlying Joplin shows no alteration or weathering that would suggest it had ever been exposed to subaerial erosion, nor are there clastic materials incorporated in the Short Creek that would suggest this. It seems that any real truncation would most probably have been submarine, performed by waves or currents in shallow water; but approximately the same effect could be produced by the interference of waves or scouring currents that could have prevented deposition in such shallow areas in the first place. The resultant break would thus be diastemic rather than erosion at a contact of disconformity. The oolite in itself is evidence of deposition in shallow, strongly agitated water (Pettijohn, 1957, p. 402–403), and it contains further evidence in the crossbedding present at different places.

In its environment of sedimentation, the oolite is much more closely related to the overlying Baxter Springs Member than to the Joplin Member. It is in effect the basal unit of the Baxter Springs, and except for its unique lithologic character, would not have been separately distinguished. Although the Baxter Springs varies widely from place to place, its different lithologic facies are everywhere conformable on the Short Creek.

In places along the Ottawa-Delaware County line nearly 2 miles east of Turkey Ford, all Boone strata above the Short Creek Oolite Member are missing and this member is overlain unconformably by the Hindsville Limestone of the Chester Series.

BAXTER SPRINGS MEMBER

The Baxter Springs Member is herein named for outcrops at the type locality in the west bluff of Spring River about 1,000 feet south of the Kansas-Oklahoma State line, 2 miles south of Baxter Springs, Kans. The member consists of three conspicuously different subunits, usually one superposed on the other in sequence but with the middle one in places changing in character, overlapping the lower, and apparently grading into it laterally. All three contain glauconite grains and phosphate pebbles indicating similarities of depositional environment. In the lower unit the glauconite is concentrated only in certain thin seams at the base, where it is sparingly and locally accompanied by phosphate pebbles. The middle unit contains glauconite irregularly, though on the whole rather abundantly, scattered throughout, and has phosphate pebbles either rather sparingly scattered throughout in some sections, or lying at or near the base of the unit where it overlaps the lower unit in other sections. In the upper unit, which is usually only a few inches thick, the glauconite is abundant, nearly diagnostic for the unit, and is commonly accompanied by small phosphate pebbles.

The three units, in sequence from bottom to top, are L, K, and J beds of Fowler and Lyden (Fowler, 1942) as most commonly revealed in superposed sequence. Our interpretation differs from those authors as to the identity of the unit that in certain areas cuts down disconformably and truncates the underlying unit or units; as indicated above, we believe this to be the middle unit, or K bed, whereas they believe it to be the upper unit, or J bed. This problem will be discussed more fully on pages 46–49 after the different units have been described. As it is necessary to designate these separate units in some way in order to discuss them, the letter designations of Fowler and Lyden, which have wide acceptance in the mining field, are used. Netzeband (1929b) has used the term "Lincolnville Chert" for L bed.

L BED

The L bed is typically a thick unit of pale chert, commonly in part cotton rock, but locally it contains a little interbedded limestone. It crops out along Spring River and its tributaries, and along the Neosho both above its junction with Spring River and for a few miles below the junction. South of the Horse Creek anticline the unit it exposed on the south point between the Neosho and Elk Rivers, and also in places along the north side of the Elk for about 3 miles to the east. Farther east, and to the south and west, it is truncated by overlap of K bed. Its south edge crosses the Neosho at the river bend 2 miles below the mouth of the Elk, curving northwest and northeast from there.

The unit is present in the major part of the mining field, but is cut out abruptly by the disconformity at the base of K bed along a very irregular northwest-southeast line that traverses the western part of the field.

Beyond this line only scattered islands of thin, and in part questionable, L bed occur. The west limit of the main mining field corresponds roughly to the edge of L bed, except in the southwestern part of the field where some very rich mines have been worked for 1–2 miles beyond its edge (pl. 3). It is probable that north of the mining field the edge of L bed may swing back to the east, for drilling on the Garrett, Burns (north of Garrett), and Thompson tracts, 3 miles north of Picher, shows L bed in some holes but not in others. Information is too scant, however, to define the edge in this area. L bed is not a very richly mineralized unit, though locally in its basal few feet it contains ore in which lead is relatively more abundant.

Thickness

The unit ranges in thickness from 0 to 60 feet. On the left side of the Neosho River, 2 miles below the mouth of the Elk River, it is 15 feet thick, but it decreases to 0 a short distance southwest of here. One mile below the mouth of Spring River it is 34 feet thick, but on the left bank of the Neosho three-quarters of a mile above the mouth of Spring River, it is only 10–12 feet thick. At Moccasin Bend the unit is 20 feet thick; on the right bank of Spring River just south of the Kansas line, it is 24 feet thick.

In the mining field, L bed as interpreted in drill cuttings may vary appreciably in thickness within a given 40-acre tract, but the variation is commonly limited to 15 feet or less. Most thicknesses are between 20 and 45 feet, with the commonest range between 25 and 40 feet. Few drill holes show as much as 50 feet, but one hole on the line between the Muncie and Black Eagle shows the maximum of 60 feet. Possibly this is a local and aberrant thickening. Maximum thicknesses (40 ft or more) are most common in a northwest-southeast belt extending 4 miles from the Lucky Jew through the heart of the mining field to the New Chicago tract, and having a width of as much as 2½ miles, measured from the southwest edge of the unit. Within areas having moderate to thick L bed, there are spots where the unit may thin unaccountably, as on the southeast quarter of the Barr and on the adjacent northwest quarter of the Blue Mound east forty, where several drill holes show an L bed thickness of only 5–12 feet.

The decrease in thickness at the southwest margin of L bed is abrupt. For example, at one place on the Robinson tract, the truncation from a thickness of 34 feet to 0 takes place within 200 feet. In the outlying island of L bed shown on the Federal Jarrett and Stanley tracts (pl. 3), the thickness reaches a maximum of 41 feet; thicknesses of 40 and 41 feet on the Stanley part are within 200 and 300 feet, respectively, of the edge of the island. The island in the Crystal-Central and Ritz shows L bed thicknesses to a maximum of 43 feet. However, in most other islands lying beyond the edge of the main L bed sheet, the thickness is somewhat less.

In two holes on the east forty of the Garrett mine tract, 3 miles north of Picher, thicknesses are 24 and 34 feet. This occurrence is also near the truncated edge of L bed.

Character

The L bed is typically a pale-gray to nearly white chert, but locally grades to light gray, light to pale blue, and light to pale brown. In places, it may contain irregular blotches, several inches across, or smaller mottling of darker gray, blue, or brown color, which are more likely to occur in the top or bottom few feet. Near the top in some sections along Spring River, concentric darker bands within the paler mass suggest enclosed nodules. The chert is massive to very thick bedded, except in the upper few feet and basal 6–10 feet in which beds are from a few inches to 18 inches thick. The thin-bedded chert layers are separated generally by thin siliceous selvage or shale partings but locally by thin limestone partings against which the chert may develop long nodular shapes. Individual beds tend to be lighter colored on the border than in the center.

The basal thin-bedded cherts are characteristically mottled by rounded, elliptic, or elongate curved spots, 1/16–⅛ inch in smallest diameter. In unmineralized areas the spots are composed of brown limestone that appears to have been recrystallized, but in mineralized ground they are dark jasperoid. The spotting on a pale-gray base suggests the mottling of a Dalmatian coach dog, and this particular type, therefore, has been termed "coach dog" mottling. (fig. 9). Although best developed

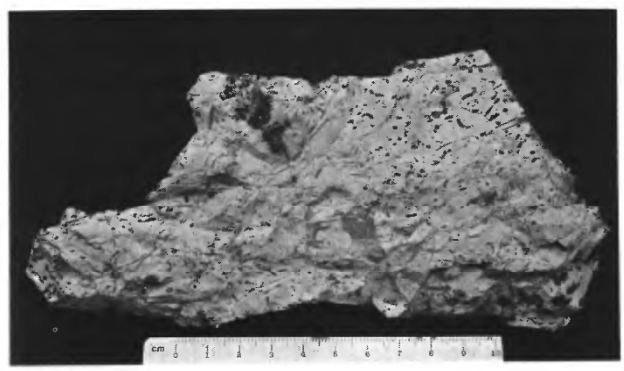


Figure 9.—Chert specimen from Consolidated No. 2 mine showing "coach dog" mottling (at left and upper right) that is characteristic of basal L bed. Specimen presented by Curtis Templain.

in some of the beds within the basal 5–10 feet of L bed, it may occur anywhere in the unit, though generally in zones only 1–2 feet thick. This mottling is also found scattered in chert of the Joplin Member but is not so abundant or conspicuous there. The resemblance is to worm borings, but the spots have the same size, shape, and pattern as cross sections of fossil bryozoan stems found in cotton-rocklike chert on the Ballard tract. These fossils are the remains of a ramose bryozoan in growth colonies and are very likely one of the two species of *Leioclema* known to occur in the Keokuk and lower part of the Warsaw (Helen Duncan, written commun., 1965). It is believed that the spots in the coach-dog mottling were formed by replacement of such bryozoan branches wherein all other traces of the organism were destroyed. The distribution is not that of detrital material, but appears to be one of growth in place, as in the bryozoan colonies

A different type of mottling occurs in the top 5–10 feet of L bed chert in many sections. This is an ill-defined gray or bluish mottling a fraction of an inch across, in linear rather than rounded shapes. It appears to be caused by finely disseminated pyrite or marcasite, and may represent a pyritic replacement of some organic material originally scattered in the siliceous mud of the sea floor.

Unlike the chert in the Joplin Member, L bed chert contains few fossils. Locally, the chert may contain a few casts of crinoid stems, bryozoans, or brachiopods, and practically all thin sections of L bed reveal sponge spicules which, however, usually constitute only a very small part of the mass.

In many sections the chert grades in part to white soft cotton rock, but the two are so intermixed that it is commonly impossible to draw a sharp distinction between them. Generally, the drill cuttings from each 5-foot cut of L bed in a cotton-rock-bearing interval contain both chert and cotton rock in varying proportions; only in a few places is an interval as much as 15 feet thick composed entirely or largely of cotton rock. The cotton rock may occur in roughly the same part of L bed in adjacent holes, but the proportions vary widely, and at the other extreme there may be no correlation whatsoever between adjacent holes. Although cotton rock may be found at any level in L bed, it tends to occur most abundantly in or near the middle of the unit, and the basal few feet of L bed are most likely to be free of it. The cotton rock may contain stylolites, which are generally of smaller length and amplitude than those in limestone.

The sparse limestone, which occurs only in some sections of L bed, tends to be associated with cotton rock and locally grades into in. The limestone is partly in blunt-ended lenses from 1 to 2 inches, rarely as much as 18 inches, thick between the cotton rock or chert beds, partly in irregular masses a few inches across and partly intermottled with the chert or cotton rock. It may appear anywhere in L bed. The limestone is fine to medium grained, brown, gray, or greenish gray grading to pale shades, and is locally somewhat glauconitic. In places it may be argillaceous.

Most sections contain glauconite in the basal 1–2 feet, in part as discrete grains in thin intervals of the chert or cotton rock, but more abundantly in one to several thin siliceous shaly partings. This occurrence of glauconite is not nearly so conspicuous as in J bed and has been commonly overlooked. In some sections the glauconite may occur sparingly in the chert, cotton rock, or limestone well above the base, or, rarely, up to the top of L bed.

Scattered occurrences of sparsely and poorly oolitic chert, in part glauconitic, have been observed near or at the base of L bed in drill cuttings and also at one underground locality on the west edge of L bed in the Gordon No. 3 mine. In the latter occurrence the oolite is closely associated with a thin zone containing phosphate pebbles. These features are of particular interest because of their bearing on the relationship between L bed and the overlapping western phase of K bed, which contains the same features in homologous stratigraphic position.

The lower part of L bed commonly has a pseudobreccia texture in which the "fragments" of nearly white opaque chert are not angular but only subangular to more or less rounded and mashed together like irregular pillows, with gray selvage lines or perhaps only vague grayish irregular lines between them. Larger fragments, an inch or more across, may be separated by a matrix made up of many similar small fragments only a fraction of an inch across, closely fitted together. The texture in cross section is most like that of the headcheese type of sausage, and is similar to a chert-replacement texture produced in the Joplin Member by redistribution of silica during leaching by ground waters (see p. 88). In places the headcheese texture may be localized along ill-defined dendritic patterns, grading into massive material. The texture usually looks accretional rather than cataclastic, but it may include two textural types that have not been differentiated. One type was produced by replacement in a muddy limestone matrix. The other was possibly produced by slumpage or slight deformation, with perhaps some solution, when the siliceous material was soft and pliant, like cotton rock. Subsequently, the whole mass has been hardened by a slight redistribution of some of the silica,

so that at present the rock breaks with a typical conchoidal flinty fracture cleanly across the fragments rather than around them.

Stratigraphic Relations

The L bed overlies the Short Creek Oolite Member conformably, and is overlain with both conformity and angular discordance by the closely related K bed.

K BED

The K bed crops out at numerous places along Spring River to its mouth, and is present below the surface over much of the mining field. It also is present in a railroad cut 1¼ miles northeast of Ogeechee, but is not present in the correlative horizon in the river bluff on the left bank of the Neosho, 2 miles farther northeast, nor on the right bank, 1 mile to the southeast. In Delaware County, K bed was formerly exposed at several places along the Neosho River and in its tributary hollows from the bend 2 miles below the mouth of the Elk River to the former bridge at Echo, and probably to the bluff east of Needmore; most of these localities, however, are now under water. The unit is present at Tynon Bluffs; on Elm Creek southwest of Grove; in the bluff at the prominent bend of the Elk River 3 miles above its mouth; and its basal bed is locally present along the Ottawa-Delaware County line nearly 2 miles east of Turkey Ford.

K bed occurs in two different facies. Along Spring River east of the mining field and in the eastern part of the field where it overlies L bed, K bed is an alternation of granular limestone and nodular or bedded chert, much like the Joplin Member except for the presence of considerable glauconite. This facies is an important ore-bearing unit. Approximately on the western fringe of L bed, K bed loses much of its nodular chert and becomes somewhat shaly, though the shale is in part very siliceous, locally resembling a soft dirty chert. The occurrences near Ogeechee and in Delaware County belong to the western facies, but they have transitional features, such as a varying, though commonly large, content of the more crystalline limestones.

Thickness

K bed varies greatly in thickness and locally may even be absent within the general area of its occurrence. Much of the extreme variation in the eastern limestone-chert facies is obviously attributable to leaching and mineralization, but there is additional variation even in unaltered sections. The thickness in some areas bears a roughly reciprocal relation to that of L bed, more conspicuously so along the western margin of L bed than farther east. The stratigraphic thickness ranges from 0 to 51 feet.

K bed is at least 26 feet thick at the top of Tynon Bluffs, and 18 feet thick in the bluff on the left side of the Neosho 2 miles below the mouth of the Elk River. The partly shaly phase of K bed is at least 24 feet thick in a railroad cut 1¼ miles northeast of Ogeechee, excluding the base which is not exposed. K bed is 4 feet thick at Moccasin Bend, and 28 feet thick on the right bank of Spring River, 1,000 feet south of the Kansas line.

Within the mining field, information on the thickness obtained from examination of K bed in mine workings in this unit was supplemented by examination of the cuttings from prospect drilling carried on during the 5 years of fieldwork within the district. As both the K bed workings and the drilling were unevenly scattered, the concept of thickness variation that emerges can be only an approximation. Although additional thousands of prospect drill holes in the field have cut K bed, the differentiation of this unit in drillers' logs is so uncertain and uneven as to have made such logs unusable for detailed thickness studies. The thickest K bed lies in two areas, one along a general zone at the west edge of L bed, the other in a flat arc circling the southeastern and eastern edges of the field. The observed thicknesses are as much as 51 feet on the Robinson mine tract in the first area, and as much as 46 feet on the Scott tract in the second area (pl. 3).

The thickness of K bed in the zone where it overlaps the western fringe of L bed (pl. 3) varies widely and abruptly between 0 and 51 feet. In all or nearly all of this area the K bed facies involved is either the completely shaly phase or one in which much shaly material is mixed with the limestone.

K bed is absent over the full thickness of L bed a few hundred feet back from the L bed margin on parts of the Ritz, Central, Gordon No. 3 northwest forty, Pelican northeast forty, Stanley, and Federal Jarrett properties; and is thin (2–7 ft) in corresponding positions on the Robinson west forty, Pelican northwest forty, and intervening properties. The nonoccurrence of K bed on the Ritz, Central, Stanley, and Federal Jarrett tracts is over relatively small outlying bodies of L bed—the Ritz-Central island, the island that lies along the State line, and the tongue that extends southeast on the Jarrett northeast block (pl. 3). As its margin is approached, L bed decreases rapidly in thickness by lowering of its upper surface, and K bed appears and increases correspondingly in thickness. The thickest K bed is near the margin of L—either just inside the margin where it overlies an attenuated wedge of L bed, or

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just beyond the margin. Thicknesses of 51 feet are attained on the Robinson, 35 feet on the Federal Jarrett, 39 feet on the Stanley (in the narrow trough between the L bed island and the main mass of L bed on the east), 24 feet on the Pelican northwest forty, 32 feet on the Gordon No. 3, 40 feet on the Kenoyer central forty, and 45 feet on the Admiralty. If a complete study could have been made of all drilling carried on along the L bed margin over the many years of prospecting in the mining field, particularly that on the Anna Beaver, a more detailed and more completely documented picture could be presented; but examinations of the cuttings from a couple of hundred scattered holes on the tracts mentioned have at least indicated a relationship that is probably valid in the unstudied stretches along the L bed fringe.

A few hundred feet beyond the L bed margin, K bed decreases in thickness, perhaps averaging between 5 and 15 feet. On the Federal Jarrett west of the L bed island along the State line, several drill holes indicate a maximum thickness of 9 feet in ground that is beyond 400 feet from the island; on the northeast side of the island, however, in the center of the 1,200-foot gap between the island and the narrow prong of L bed shown on plate 3, the thickness is 19–25 feet. On the Laura Jenny Zheka tract, the thickness is 3–15 feet beyond a distance of 1,000 feet from this same State line island of L bed. On the John Beaver northeast forty, which lies partly behind a big island of L bed on the Anna Beaver, the thickness is 5–27 feet but mostly 12–20 feet, with the thickest occurrences about a quater of a mile beyond the main edge of L bed. Extensive drilling on the Blue Goose No. 2 tract (80 acres) shows thicknesses mostly of 5–7 feet, rarely exceeding 10 feet, but in several holes, K bed is either absent or unrecognizable. Some of these absences are associated with a small thin island of questionable L bed. Other areas of very thin shaly facies of K bed beyond the L bed margin include the Blue Goose No. 1 (1–5 ft), Wilson (0–3½ ft), and Scammon Hill (0–6 ft). These thin sections are all significant in that, except for about 1 foot of J bed, they represent the complete thickness of the Baxter Springs Member, in contrast to thicknesses of 70–80 feet in the eastern part of the field (Scott and Crane tracts) and 65 feet on the Bendelari.

An anomalous thickening of K bed, as compared to other areas west of the L bed margin, extends in a narrow belt southeast of the Ritz-Central island of L bed (pl. 3). The maximum thickness observed, 36 feet, was in a drill hole on the Ritz-Jay Bird line a few hundred feet from the island. The belt crosses the southwest part of the Woodchuck and extends onto the northeast forty of the See Sah. Although further extension of the belt has not been accurately determined, it probably swings east to join the thick K bed zone on the Admiralty. This belt of thick K bed occupies a position analogous to that of a similar belt on the John Beaver that lies southeast of the large island of L bed on the Anna Beaver tract.

In the mineralized area on both sides of the State line southwest of Melrose, Kans., K bed is 3–24 feet thick in unaltered sections.

K bed is present to the exclusion of L bed in several drill holes, cuttings of which were examined on the west forty of the Garrett mine tract, 3 miles north of Picher, where it is 9–19 feet thick. A couple of holes on the east forty (NE¼NW¼ sec. 36) contain L bed but no overlying K bed. The same complementary relationship between the occurrence of L bed and K bed is found on the Thompson tract (NE¼SE¼ sec. 36) and Burns tract (SW¼ sec. 25, T. 34 S., R. 23 E.); thicknesses of K bed are 8–9 feet and 14–18 feet, respectively. The drilling examined was not complete enough to define the edge of L bed and its relationship to thicknesses of K bed in this general area.

Reciprocal relationship between thicknesses of K and L beds along the L bed margin in Delaware County is indicated in sections along the left bank of the Neosho River below the mouth of the Elk River. On the point 1 mile southeast of the mouth of the Elk, at least 21 feet feet of L bed is preserved, but higher strata are eroded; hence, the former presence or absence of K bed cannot now be determined. To the southwest 1¼ miles, 4 feet of K bed overlies 15 feet of L bed; but only a few hundred feet farther southwest, 18 feet of K bed overlaps the edge of L bed and rests directly on the Short Creek Oolite Member. At the last locality, the oolite is now a few feet below water at the usual levels of Lake O' the Cherokees.

Character of the Eastern Phase

The eastern phase of K bed is a coarse crinoidal to rather fine-grained brown to gray limestone, variably glauconitic, with interbeds and nodules of chert. The chert beds range in thickness from a few inches to 4 feet, and the nodules are as much as 9 inches thick. In unaltered and uncondensed sections, the chert forms less than half, and in places, as little as 10 percent. It is gray, brown, or blue, grading to pale shades or locally to dark shades; these colors may be irregularly intermottled, or the nodules may be concentrically banded. As the several colors commonly appear in the same section, in drill cuttings this unit is usually the most variegated in the Boone. Much of the chert is glauconitic, as are also thin

shaly partings that are probably residual from solution of the limestone.

The chert contains silicified fossils, including cup corals and abundant remains of small fossils, some of which are sponge spicules. Chert in the lower part may also have a gray linear mottling similar to that in upper L bed; this mottling is apparently derived from replacement of some organic material by finely disseminated pyrite or marcasite, though details of the organism are not preserved.

In many sections of this facies of K bed, the upper part, which may be more or less than the upper half, contains less chert and more glauconite in both limestone and chert than the lower part. The chert, also, tends to be more deeply colored.

A few sections contain cotton rock which may be mottled rather intimately in the limestone, or in a more advanced stage, may replace masses of the limestone except for coarse calcite grains that are the remnants of crinoid stems. The cotton rock shows the usual gradations to pale chert.

Small dark-brown phosphate pebbles, generally ¼–¾ inch in diameter, are scattered in the limestone and may be more common than recorded observations would indicate, as the pebbles are not very conspicuous.

Drill holes on the Blue Mound tract and on the Chubb west forty have cut oolitic chert in K bed. This oolite is less perfect and more diffuse than that in the Short Creek, and it differs further in being highly glauconitic. In both properties it is above L bed chert, which is 30 feet thick in the Chubb section. The oolite on the Blue Mound is apparently scattered throughout a 6- or 7-foot total section of K, whereas on the Chubb, it is in the basal third of a 19-foot section.

Character of the Western Phase

Drilling concurrent with the period of fieldwork was not compete enough to define exactly the narrow belt of transition between the eastern and western phases of K bed, but this belt is not far east of the L bed margin. The eastmost observance of shaly material was 2 miles from the margin, in a 5-foot zone at the middle of 34 feet of K bed, at the base of the less cherty upper part. This shaly material was in a drill hole on the Pat, though it was not recognized in four other holes on that tract. A little fine-grained siliceous shaly material shows in a few holes on the New Chicago tract where it may appear at the top or bottom of K bed. The facies also forms part of the thin K bed unit on the Cherokee and it forms the upper half of a 9-foot section of K bed on the Kansouri. The eastern facies of K bed is present in the eastern half of the Tulsa Quapaw mine, but several drill holes in the western half show the shaly western facies, 4–12 feet thick, overlying comparable thicknesses of the eastern facies. In the adjacent part of the Robinson mine, the two facies are similarly superposed, but farther west on the Robinson, only the thick shaly phase is represented. On the Admiralty, the transition occurs within half a mile between the northeast corner of the north forty and the southwest corner of the 20-acre tract next on the south; the edge of L bed is in the same gap. The shaly facies takes the place of L bed near its margin on the Garrett, Thompson, and Burns tracts, 3 miles north of Picher.

The shaly western phase is typically a thin-bedded siliceous earthy limestone, dark-gray or dark-brown, but the colors may grade to light and pale shades. Greenish gray is also a common color. The rock is fine grained and moderately to sparsely glauconitic. In many places it is thickly and finely spangled by amorphous pyrite. Characteristic of the facies is the abundance of bryozoan remains, chiefly as the imprints of fronds on the bedding planes but also as stemlike parts. The limestone may also contain abundant siliceous sponge spicules. Scattered crinoidal calcite grains are locally common, and in places so increase in abundance that parts or the whole unit become a coarse-grained limestone that is more glauconitic than the rest, and brown to gray, grading to lighter shades. The fine- and coarse-grained material may be interlayered, or one may be below the other, the coarse-grained probably occurring below in most localities. Where the whole unit is coarse grained, it is indistinguishable from the less cherty upper part of the eastern facies of K bed.

Chert is commonly absent in the western phase, or subordinate, perhaps 10–20 percent, but locally forms more than half the unit. It may be finely marbled as amoeboid blebs in the limestone or segregated in larger nodular masses or bedded streaks. It is like the chert in the eastern K bed facies, particularly that in the upper part, and is conspicuously mottled by fine organic remnants, including sponge spicules.

The lower part, or locally all, of the western facies in many places also contains glauconitic oolite, partly in chert, like that described in the eastern facies. It is far less perfect and texturally more diffused than the Short Creek oolite. Where this phase of K bed is superposed directly on the Short Creek, there may be a question as to whether the oolite should be all classed as Short Creek. However, there can be no question where the two are separated by L bed, as along the west line of the Bendelari, on the Tulsa Quapaw, on much of the Robinson, and on part of the northwest forty of the

Anna Beaver. The K bed oolite shows in many holes in the mineralized area southwest of Melrose, Kans.

At or near its base, the western facies locally contains a thin zone of phosphate pebbles closely associated with the glauconitic oolite. These pebbles are rounded to elongate, as much as half an inch in diameter, and are in a matrix of glauconitic limestone that may range from fine to coarse grained. Where observed underground along the north side of the Tulsa Quapaw mine, in the eastern part of the Robinson mine, and in the northeastern part of the Midcontinent mine, the pebble bed is as much as 1 foot thick, and overlies a thin remnant (1–10ft) of the eastern facies of K bed. In the Anna Beaver mine near the south line of the northeast forty, the phosphate pebble zone, 1–2 inches thick, is just above the thinned wedge of L bed. A quarter of a mile to the southwest, it is at the base of 20 feet or more of the western facies and above a thin mineralized interval of the Joplin Member, but remnants of Short Creek oolitic chert intervene in places. At both of these Anna Beaver localities the phosphate pebbles are accompanied by fish teeth. Elsewhere in the Anna Beaver the phosphate pebbles occupy these same positions, or may lie near, but not actually at, the base of the western facies. However, they may be entirely wanting. In the Gordon No. 3 mine at the edge of L bed, two phosphate pebble zones appear in the lower part of K bed. The lower, 1½ feet thick, lies 3–4 feet above the base of K bed, and the upper thinner one lies 3½ feet higher.

The western phase has not been observed widely on the surface, but a predominately shaly section with coarse-grained limestone confined largely to the lower quarter is present in the railroad cut 1¼ miles northeast of Ogeechee; a similar section, with the coarse-grained material in the upper half, however, was formerly exposed in the left bluff of the Neosho, 2 miles below the mouth of the Elk River. Former outcrops on the west side of the Neosho just south of the Delaware County line contained both oolite and phosphate pebbles near or at the base of the unit, and those in the bluff east of Needmore contained phosphate pebbles at the base. Strata of the western phase exposed at Tynon Bluffs contain phosphate pebbles in glauconitic limestone of the basal 6–8 inches, but the pebble zone here is not persistent along the outcrop.

Stratigraphic Relations

The eastern facies of K bed overlies L bed conformably. The western shaly facies, however, bears an anomalous relation to L bed, not easily ascertained because exposures are lacking at most critical points.

One to five hundred feet inside the western margins of L bed, including those of the outlying "islands" in the mining field, K bed is lacking in many places over full thicknesses (35–43 ft) of L bed, whereas at or a short distance beyond the margins, maximum thicknesses of K bed are present, with no L bed. Drill holes in the intervening areas show K bed appearing as a thin wedge over L bed and increasing regularly in thickness as L bed decreases. The two units are sharply distinct. There would thus seem to be a simple disconformable relationship between them. The contact lacks any resemblance to an erosional unconformity. It seems more probable that the L bed masses were biohermal or biostromal bodies, initially calcareous, growing on the floor of the shoaling sea near the beginning of the Meramec Epoch; and that the more shaly K bed facies was spread by marine currents as a sediment from the west against the edge of these growths, lapping up on their flanks in places, filling the irregularities of their sloping faces, and accumulating to maximum depth in the more protected zones along the foot of the slopes or in the lee of outlying biohermal islands (pl. 3).

From the known prominence of algae as reef makers, it would seem possible that much of the original body making up these bioherms and biostromes may have been of algal origin. However, there are no recognizable fossil remains of the type commonly described in substantiation of such origin. Some of the headcheese texture of apparent replacement origin in lower L bed (p. 42) might possibly represent a texture produced by algal growth with segregation of shaly clastic material between the growth heads. If so, later silicification has obliterated the organic details. Perhaps this is to be expected. Howe (1932, p. 60), in discussing the role of blue-green algae as precipitants of ancient limestones, says,

> Why should one expect their delicate structure to persist for millions of years? Nevertheless, one who is accustomed to see and to handle the algae of the present day may feel convinced from their macroscopic characters that certain laminated ancient limestones were laid down by algae, even while admitting more or less subconsciously the possibility of being deceived.

Goldring (1938, p. 21–31, 59) has emphasized the common association of algal reefs (or biostromes as described) with oolites. The association is apparently attributable to a shallow-water environment favorable for development of both.

Probably the macrofauna composing these reef bodies was largely bryozoan and sponges, rather than crinoidal. Not only are bryozoan imprints the most abundant fossils in marginal zones of the L bed chert, but if the interpretation of coach-dog mottling as vestigial bryozoan stems growing virtually in place is correct, the abundance of this mottling in L bed is additional evi-

dence of bryozoan prominence. Their abundance as fossils in the shaly phase of K bed would also be expectable. Silicification to the characteristic L bed chert and cotton rock came shortly after the initial deposition, and was comparable to the chertification of the crinoidal bioherm in the Reeds Spring Member, previously described (p. 24).

Several exposures in the mines corroborate the stratigraphic relations between L bed and the shaly phase of K bed, as outlined above, and give further information on the mutual relations of these two units. A revealing exposure in the Kenoyer mine (SW¼SW¼ sec. 20, T. 29 N., R. 23 E.) shows the wedge end of L bed chert decreasing in thickness from 7 feet to 0 in a horizontal distance of 10 feet, with the shaly facies of K bed on top, bedded parallel to the contact (fig. 10). Thus, the bedding in K bed here partakes of the same dip as the upper surface of the L bed wedge. The relationship is obscured somewhat by intense leaching that has here affected the underlying Joplin Member long subsequent to deposition, but it resembles the draping of finer grained limestone beds over the crinoidal bioherms in the lower Reeds Spring. In the extension of some of the lower limestone beds, however, there is apparently abrupt gradation along the bedding from the siliceous limestone to massive chert, as though the limestone had been replaced laterally by chert.

Exposures in the Mehunka Zheka Beaver (Bird Dog 14) mine show the same type of discordance, in which the shaly facies of K bed overlies an abrupt wedging out of L bed and is stratified parallel to the dipping upper surface of the wedge.

The preceding examples have emphasized that at least much of the shaly phase of K bed is younger than L bed. Other exposures, however, indicate contemporaniety of the two, even at the stratigraphic level of basal L bed. Intertonguing and replacement of the siliceous shaly facies of K bed by chert of the "coach dog" zone in the lower part of L bed, only a few feet from the L bed margin, is well shown in the Anna Beaver (fig. 11), and Gordon No. 3 mines. The chert along the contact between the two tongues shows the headcheese texture. This texture is elsewhere very common in the thin wedge of chert that marks the extreme limits of basal L bed, and apparently is diagnostic of replacement in fine-grained siliceous limestone at this horizon. However, other patterns of replacement may prevail locally. The siliceous shale partings that commonly appear between the beds of lower L, well away from the L bed margin, are very similar to the shaly phase of K bed and might be viewed as the extreme example of intertonguing between L and K bed lithologies.

Stratification of K bed is not always parallel to the underlying L bed contact on the sloping L bed margin, for, in at least two places, higher strata of K bed were seen to overlap lower ones onto the flank of L bed. One locality is in the Anna Beaver mine (NW¼NE¼, sec. 19). Here, the overlapping strata of K bed belong to the coarse-grained limestone phase of the western facies; their stratification is practically parallel to that in L bed (fig. 12). The other locality is in a cut of the St. Louis-San Francisco Railway, about 1¼ miles northeast of Ogeechee where a similar lithology exists.

Although the lapping of K bed over L bed with parallelism of the bedding suggests lateral gradation between them, lack of continuity in this bedding at most places leaves some doubt. However, at one locality on the edge of L bed on the Gordon No. 3 tract, one of the two phosphate pebble zones here present near the base of K bed continues laterally for some distance along a parting in the "coach-dog" zone at the base of L bed; and glauconitic oolitic chert that is closely associated is incorporated into lower L bed. Elsewhere, the marginal part of L bed where it is overlapped by K is usually browner and more earthy than typical, and contains abundant bryozoan casts. Thus, the marginal zone shows gradations toward and into the K bed type of lithology, and at the same time, it is overlapped by K bed with both accordance and discordance in bedding. It is apparent that the lower part of K bed west of the L bed margin is equivalent in age to L bed farther east. The shaly material was being deposited in deeper quiet water to the west while the biostromal material, which was soon silicified to L bed, was growing in shallower water to the east. Greater agitation in these shallow waters above effective wave base prevented settling of the shaly material, except intermittently as partings. There was some intertonguing of the two types of material in the early stages. In places, they accumulated at approximately equal rates to produce parallel bedding; elsewhere, the biostromal material accumulated more rapidly and developed steep reeflike slopes along its western margin which were later mantled by the shaly facies, thus producing the discordances in bedding previously described. The supply of mud contributing to the shaly facies was not large, and marine currents were such that more of it was trapped near the protecting buttress of the L bed margin with its outlying "islands" than in the unobstructed basin a short distance beyond. Eventually, conditions changed eastward so that over much of the mining field, L bed was succeeded by the eastern facies of K bed consisting of limestone and chert. Farther west, however, the shaly facies in places persisted through all K bed deposition.

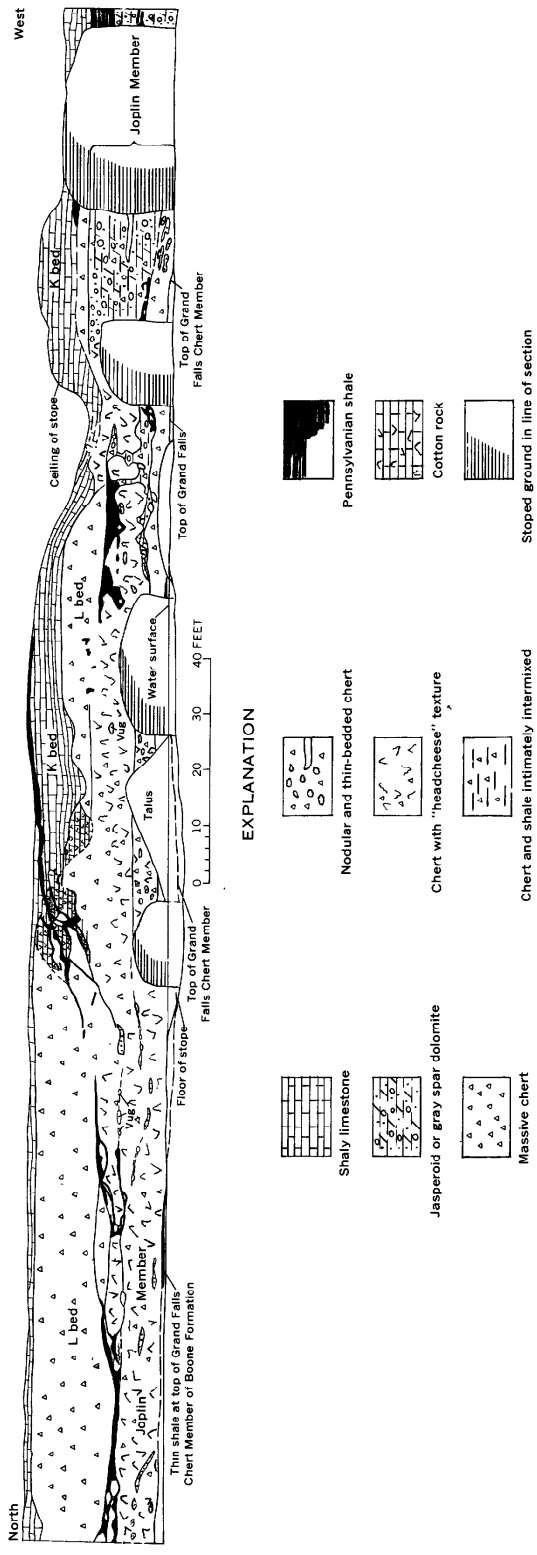

FIGURE 10.—Section showing western margin of L bed overlapped by shaly facies of K bed in Kenoyer mine. Stratigraphic relations somewhat obscured by intense leaching of Joplin Member of the Boone Formation, some slumpage, and introduction of Pennsylvanian shale. Section curves from nearly north-south at left end to east-west at right end.

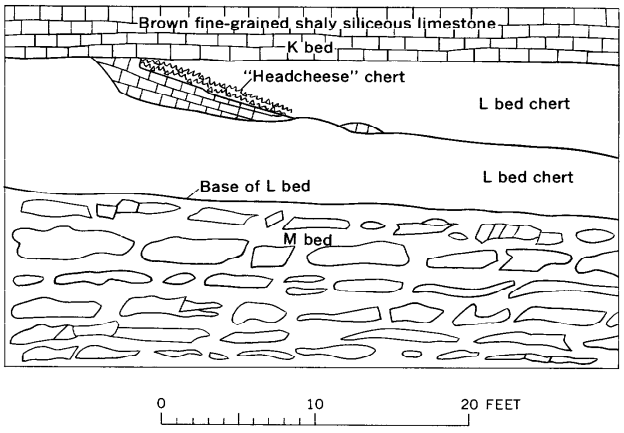

FIGURE 11.—Diagrammatic sketch showing intertonguing of the shaly siliceous limestone phase of K bed with chert near the base of L bed, in Anna Beaver mine at locality near center of northeast forty acre tract. Margin of L bed 20 feet to left.

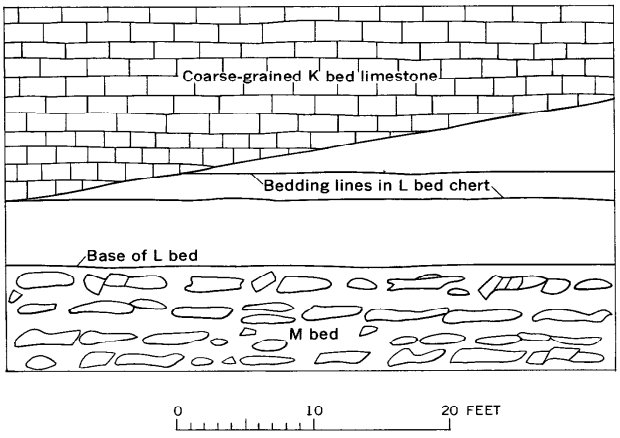

FIGURE 12.—Diagrammatic sketch showing relation of bedding in coarse-grained K bed limestone to the thinning wedge of L bed near its margin in Anna Beaver mine.

Where K bed laps beyond the margin of L bed, its base is the time equivalent of basal L bed, and it is conformable on the underlying Short Creek Oolite Member.

J BED

As herein interpreted, J bed is a highly glauconitic coarse-grained to shaly fine-grained limestone, generally 1 foot or less but locally 5 feet thick, and characteristically containing at its base numerous small dark-brown phosphate pebbles as much as 1 inch across. These pebbles are well rounded and polished; they may enclose grains or streaks of glauconite. Commonly, the pebbles are accompanied by scattered platelike fish teeth. In a few places, additional zones of phosphate pebbles may occur as much as 3 feet above the base of J bed. The basal layers of J bed and the underlying 2 inches of K bed are in many places traversed by circular tubes, a quarter of an inch in diameter, filled with more glauconitic material and rarely a small phosphate pebble. These tubes are usually revealed in circular or elliptic cross section. They record the borings of some organism.

J bed contains chert of several colors similar to that in K bed, but in addition, a very dark gray or dark-brown, almost black chert is conspicuous in many places; the chert, especially the lighter colored varieties, have a greenish tinge in a few places. The masses are very irregular in shape. As in K bed, much of the chert is glauconitic and contains remnants of small organic materials, including sponge spicules.

J bed marks the culmination of glauconite deposition in the Baxter Springs Member and is at the same time the most highly glauconitic unit in the stratigraphic sequence. Altough not universally present, J bed is conspicuous in most sections. It resembles the more crystalline glauconitic limestone in the western facies of K bed, especially where such limestone contains phosphate pebbles and fish teeth. Fowler and Lyden (in Moore, R. C., and others, 1939; Fowler, 1942) have, in consequence, interpreted the western facies of K bed as simply a local thickening of J bed, and have applied the J bed designation to it. However, the continuance of the thin phosphate-pebble zone from eastern areas over the top of the thickened glauconitic unit in western areas makes desirable the simple direct correlation of this restricted unit as J bed, although it appears that conditions of sedimentation were identical in the two units, simply starting at a slightly earlier stage in the western area.

J bed is conformable on K bed or on L bed where K is absent. Where both K and L are absent, as in some areas on the southwest fringe of the mining field, J bed rests conformably on the Short Creek Oolite Member, a relationship that is interpreted from drill-hole cuttings, for we have seen no underground exposures showing it. Where the Baxter Springs Member is represented by only a foot or less of glauconitic limestone, it is impossible to distinguish, in drill cuttings, J bed from coarser grained limestone phases of K bed; but the persistence and uniformity of J bed at this horizon makes reasonable the assignment of such material to J bed.

CORRELATION AND AGE OF BAXTER SPRINGS MEMBER

The coarser grained limestones in both the eastern and western facies of K bed and in J bed contain an abundant and varied fauna. Most fossil collections

studied have been from the eastern facies of K bed. In the shaly western facies, bryozoans are the dominant fossil remains, although scattered fish teeth and, rarely, a trilobite are found. Gordon reports (written commun., 1965) as follows on the faunal content of the member:

> The fauna of the Baxter Springs Member, as represented by 21 collections from the Wyandotte quadrangle, consists of nearly 100 species of fossils, including corals, bryozoans, brachiopods, crinoids, blastoids, mollusks, trilobites, and fish remains. Twenty of these collections came from K bed, which has one of the most prolific faunas in the pre-Chester rocks of the Tri-State mining region. One collection from J bed shows that this limestone bed with phosphatic pebbles is closer faunally to K bed than to any other part of the Mississippian section. Thus, there is no faunal evidence for a marked hiatus beneath J bed as postulated by some geologists.
>
> As at other levels in the Boone Formation, the brachiopods predominate. Continuing species that appeared first in the Grand Falls Member and range upward into the Baxter Springs Member include: *Orthotetes keokuk* (Hall), *Avonia williamsiana* Girty, **Marginatia? crawforsvillensis* (Weller), **M? mesialis* (Hall), *Labriproductus wortheni* (Hall), *Echinoconchus biseriatus* (Hall), *E. vittatus* (Hall), *Rhynchopora palmeri* Girty, *Spirifer rostellatus* Hall, *S. tenuicostatus* Hall, *Brachythyris suborbicularis* (Hall), *Pseudosyrinx keokuk* Weller var., **Torynifer pseudolincatus* (Hall), and *Eumetria verneuiliana* (Hall). Those species marked with an asterik (*) in the list are not known above the Baxter Springs Member.
>
> Typical Warsaw brachiopods in the member include *Perditocardinia* n. sp., *Marginatia* n. sp., *Marginirugus magnus* (Meek and Worthen). *Spirifer subaequalis* Hall, *S. washingtonensis* Weller, *Syringothyris solidirostris* Weller, and *Cranaena sulcata* Weller.
>
> Throughout much of K bed, *Marginirugus magnus* is represented by a small variety; but at some localities, mostly in its upper part, are specimens that match in size the typical large examples found in the Short Creek Member and in J bed. In the type section of the Baxter Springs Member on Spring River, K bed limestone 7–10 feet below the base of J bed is crowded with a small species of "nutmeg coral," *Hadrophyllum* aff. *H. tennesseense* Miller and Gurley.
>
> The abundance of *Marginirugus magnus* in the Baxter Springs Member and its virtual restriction to this member and the underlying Short Creek Member make it convenient to regard these beds as constituting the *Marginirugus magnus* zone. Beds with *Marginirugus magnus*, exposed over a considerable area along the Mississippi River in southwestern Illinois, were formerly included in the Keokuk by most workers, but more recently, particularly since the work of Van Tuyl (1925, p. 184–203), these beds and the presumedly equivalent geodiferous shales in the type section at Warsaw, Ill., have been included in the Warsaw. The Baxter Springs Member, therefore, appears to be at least partially equivalent to the lower part of the type Warsaw section.

Huffman (1958, p. 50–51) has suggested the correlation of J bed of the Ottawa County section (=K and J beds of this report) with the Tahlequah Member of the Moorefield Formation as defined by him in the general vicinity of Tahlequah, Okla.

MOCCASIN BEND MEMBER

The Moccasin Bend Member, herein named for the locality on Spring River 6 miles east of Miami, is the uppermost unit of the Boone, and is composed of alternating chert, cotton rock, and fine- to medium-grained limestone. It is exposed along the Neosho River below Miami to the mouth of Spring River, along Spring River north to the Kansas line, and along the tributaries of Spring River from the east, including Lost Creek. There are also exposures along the Neosho River in Delaware County south of the Horse Creek anticline, though some former exposures in this area are now below lake level. The distribution is imperfectly known, for the members of the Boone have not been mapped in detail for the present report.

At most outcrops only part of the member is exposed, most commonly the lower part. The top is generally obscured in soil-covered flats or benches at the tops of bluffs. Complete or nearly complete sections are exposed in the bluff on the east side of Spring River, 3 miles southeast of Moccasin Bend, in the caved pit over the Discard mine, and in underground sections in the Roanoke, Wilson, and Crystal mines. Prospect drilling in the mining field has also furnished abundant information on the unit in this general area.

The Moccasin Bend Member comprises the lettered units from B to H beds in the section of Fowler and Lyden (Fowler, 1942). The tripoli beds near Seneca are in the upper part of the member.

THICKNESS

Within the Wyandotte quadrangle, the Moccasin Bend Member is 0–140 feet thick. Like other limestone-bearing units in the Boone, this member has a great variation in thickness that is due to underground solution and collapse, either within itself or within underlying units. Where collapse is in underlying units, the effect on the Moccasin Bend Member may well be an increase in its thickness. In addition to this obviously secondary factor that influences thickness, prospect drilling in the mining field reveals considerable variation in thickness of the unit over individual 40-acre tracts, even within ground that is apparently unaltered. This variation may in extreme cases amount to 33 percent of the maximum thickness, but it is usually less than 20 percent.

When these detailed local variations are discounted, the Moccasin Bend Member is found to increase rather gradually in original stratigraphic thickness from the eastern to the western part of the mining field. This in-

crease is from about 60 feet near Spring River to 140 feet in the mineralized area southwest of Melrose, Kans.

In the Discard mine cave-in at the east end of the mining field, the member is 64 feet thick. Thicknesses of 50–65 feet were found in unaltered ground in practically all drill holes that were examined in the area from here west to the longitude of the Dardene-Maxine-Pat tracts. To the northwest, the thickness has increased to 70 feet in the single hole examined on the Coe land that lies just south of the Paxson mine tract.

Thicknesses of 65–80 feet prevail on the Harris tract in Kansas and the Dardene, Maxine, Pat, Bullfrog, and John Hunt tracts in Oklahoma, although the Maxine, with one of the largest variations in the field, also shows thicknesses as low as 55 feet.

West of these properties, the unit increases in thickness, but the extremes of variation on individual tracts overlap in such a way that the increase cannot be simply zoned or contoured. In scattered properties that lie in one roughly north-south zone across the field, from the Admiralty through the John Beaver, Pelican, King Brand, and Robinson to the Garrett, thicknesses are 66–110 feet. East of this zone, thicknesses of 100 feet are attained only in the thickest sections on the Barr property, whereas west of it, most tracts have thicknesses that attain or exceed this figure. Maximum thicknesses of 108–121 feet within the main mining field are found on the northwest, southwest, and southeast forties of the See Sah tract. Probably a zone of near-maximum thickness extends from here to the Scammon Hill tract, which shows thicknesses of 92–114 feet.

West of the main field, the thickness is 113–126 feet in holes drilled in unaltered ground on the T. R. Smith tract on Fourmile Creek (SW ¼ SW ¼ sec. 16, Oklahoma), and 99–140 feet in the mineralized area along the State line southwest of Melrose.

In surface exposures southeast and south of the mining field, the thickness is about 55 feet on Spring River southeast of Lincolnville. It increases to about 75 feet in the bluff on the left bank of Spring River 3 miles southeast of Moccasin Bend. At the type locality, 71 feet of the unit is exposed to the top of the bluff; this is probably near the full thickness. Along the Neosho River in Delaware County the thickness has apparently decreased somewhat. Although no complete section was found, partial sections formerly exposed northwest of Grove, which in composite are believed to comprise all the Moccasin Bend, indicate a thickness of about 65 feet. The member is truncated to zero thickness by the unconformity at the base of the Hindsville Limestone at one locality along the Ottawa-Delaware County line east of Turkey Ford, and it is possible that this relation may exist elsewhere in the southeastern part of the quadrangle.

Character

Although the rocks composing the Moccasin Bend show a certain similarity of the limestones and cherts throughout the total thickness, there are minor differences between different parts of the unit that can be only or best recognized in the unaltered sections. However, the distinctions are not sharp or may fail altogether, and boundaries are commonly gradational; hence, clear-cut subdivisions are lacking. Even where units can be correlated between different sections, their thicknesses vary widely. In general, there is a lower part, comprising the G and H beds of Fowler and Lyden (Fowler, 1942), composed of chert and fine-grained limestone, and an upper part, comprising the B–F beds of Fowler and Lyden, in which cotton rock is more or less prominent in addition to chert and limestone, and in which the limestone tends to be slightly less fine grained and slightly lighter colored. In the western part of the mining field, this upper part has at its top a fairly distinctive somewhat coarser grained limestone unit, B bed, in which chert and cotton rock, though irregularly present, are on the whole subordinate.

G and H beds

No lithologic distinction has been recognized between G and H beds, but because mineralization is limited to the upper (G bed) in some parts of the mining field and to the lower (H bed) in others, the distinction was made by Fowler and Lyden (1932) for reports on mining geology. The two beds usually aggregate 25–40 feet in combined thickness. On the Vanatta and A. T. Wright tracts southwest of Melrose, their combined thickness is somewhat greater than that prevailing in other areas, including other mineralized tracts along the State line southwest of Melrose; judging from a very few holes drilled, the thickness appears to be about 55 feet.

Typically, the G and H bed part of the Moccasin Bend shows an alternation of fine-grained, almost dense, limestone with hard semitranslucent or waxy-appearing chert in beds from 1 to 8 inches thick, an occasional bed reaching a maximum of about 15 inches. The chert may also occur in certain beds as nodules or less regularly shaped masses, in part as small and very irregular amoeboid blotches that are mottled in the limestone; or equally irregular blotches of limestone may occur as residuals in the chert beds. In some layers the limestone and chert are intricately intermottled in about equal proportions. Although the boundary between

chert and limestone may be very irregular and interlocking, it is generally sharply defined. Beds showing the more intricate intermixtures weather to a very hackly outcrop. In the most unaltered sections the chert is somewhat more abundant (50–70 percent) than the limestone, or the two are about equal; but in a few sections the chert averages only 40 percent. Leaching of the limestone in altered ground commonly increases the percentage of chert.

G and H beds are well colored over much of the mining field. The limestone is a full brown or locally, dark reddish brown, and the chert is various shades of brown, blue, and gray, in part with darker colors finely mottled in the lighter ones. Some of this mottling resembles the coach-dog pattern of basal L bed but is smaller and less perfect. The blue and brown chert may occur together or one may dominate the other. The rock colors are darkest in the lower few feet, but fade irregularly upward; hence the top of the G–H bed zone, as judged in drill cuttings, may be indefinite. The colors may also be less noticeable or virtually absent in parts of the mining field where much of the chert is dominantly light to pale gray and the limestone is in light to pale tints of brown and gray, more characteristic of higher zones of the Moccasin Bend. Locally, the limestone through intervals of a few feet may be shaly, or have a greenish-gray tinge; such material has little or no accompanying chert.

The cherts of G and H beds contain sponge spicules or less regular types of fossil detritus that may be derived from sponges. Sparse glauconite is present in the limestones in many scattered areas, and is most likely to occur near the base.

Although cotton rock is not common in the G–H bed zone, there are places where it is locally prominent, as on parts of the northeast forty acres of the See Sah tract and on the Bullfrog. The associated chert in such places is pale colored.

In parts of the Crane and St. Louis 4 tracts in the eastern part of the mining field, the limestone of the G–H bed zone contains medium-grained crinoidal fragments in a thin zone about 15 feet above the base. A similar thin zone of medium-grained limestone occurs 20 feet above the base in the section at the type locality of the Moccasin Bend. Scattered thin lenses containing crinoidal material, but of small lateral extent, have been noted elsewhere at different stratigraphic levels, but are, on the whole, rare.

G and H beds are the part of the Moccasin Bend most commonly seen in outcrop. A well-known exposure is at the Devils Promenade in the bluff of Spring River southeast of Lincolnville. The "promenade" is a rock ledge a few feet wide near the foot of the bluff, paved with fist-sized chert nodules or "biscuits" on the upper surface of the containing bed. J bed lies 5 feet below the promenade.

In its general lithologic aspect, the G–H bed part of the Moccasin Bend locally resembles phases of the Reeds Spring Member. Along with some underlying strata, it was confused with Reeds Spring at one locality west of Wyandotte by Laudon (1939, fig. 1), and the lower lying Short Creek Oolite Member was, in consequence, erroneously ascribed to the St Joe Member.

HIGHER BEDS OF MOCCASIN BEND MEMBER

In higher parts of the Moccasin Bend, the limestone is a paler brown or light to pale gray to almost white, and is in part somewhat more crystalline, grading from fine to medium grained in some beds. The cherts are characteristically opaque or semitranslucent, light to pale gray or almost white, but with blue to pale blue or brown tints in many sections. Although the blue tints have stratigraphic continuity in some areas, when considered more broadly, they "come and go" so erratically, both vertically and areally, that we have not been able to establish stratigraphic significance with any confidence. The upper part of the Moccasin Bend contains cotton rock, generally grading into the paler chert which takes its place completely in some sections. Except in two recognizable zones in the mining field, the bedding tends to be more massive than in G and H beds, and commonly limestone, cotton rock, and chert may be irregularly intermixed on a rather fine scale or gradational into each other over considerable thicknesses, the relative proportions varying widely and abruptly from place to place. Some of the more presistent cotton rock zones, particularly the D and F beds, are recognized in the stratigraphic classification of Fowler and Lyden (Fowler, 1942). Even within the limits of the mining field, however, the recognition and correlation of these cotton-rock-bearing units above the G–H bed zone is impossible in many places. Either the cotton rock beds give way laterally to limestone and chert like other units in the upper part of the Moccasin Bend, or cotton rock may appear irregularly in parts of the section that over most of the area are free of it, even in the best characterized and most persistent colored unit, G and H beds, at the base of the Moccasin Bend. In some sections observed in drill cuttings the Moccasin Bend is alternating pale chert and pale limestone, practically indivisible from bottom to top except for somewhat thicker beds in the upper part and an average coarsening of grain size upward in the limestones. An independent complication in mineralized sections is the usual conversion of cotton rock to pale chert—practically indis-

tinguishable from the normal chert—caused by permeation of silica from the ore-bearing solutions. Thus, a mixture of cotton rock and limestone is converted to chert and jasperiod.

The best defined unit above the basal G–H zone is the E bed of Fowler and Lyden. Although part of the limestone in this unit tends to be medium grained, grading locally to rather coarse grained crinoidal, in many places it differs in no appreciable way from the rather fine-grained phase that characterizes much of the Moccasin Bend. The associated chert, where examined in underground sections in the western part of the mining field, is interbedded with the limestone in beds 2–12 inches thick, in long regular nodules, or is in the limestone in small irregular masses that may contain irregular residuals of the limestone, or the two may be intricately intermottled on a fine scale. In addition to the usual light to pale-gray chert, some is gray or blue to light blue, the darker color in part mottled in the lighter, or zoned in the center or near the periphery of the chert masses; a few chert masses have brownish centers. The chert is irregularly distributed, and in certain drill holes as much as 10 feet of limestone may be nearly free of it. Underground sections have parting films of greenish-gray or gray shale at different levels in the limestone. Drill cuttings from the Federal Jarrett, Robinson, and Bendelari tracts in the northwestern part of the mining field and from the John Hunt tract, 2 miles west of Quapaw, have a little glauconite, part of it unusually pale colored, in coarse-grained limestone at a level interpreted as part of E bed.

In underground exposures, E bed can usually be differentiated from the more massive and generally more cherty strata lying above and below. In sections studied in Roanoke, Wilson, Crystal, and Discard mines, it ranges in thickness from 8 to 21 feet and lies somewhere in the interval 32–53 feet above the base of J bed (pl. 4). But E bed was not recognized in the lower 50 feet of Moccasin Bend as exposed in the East Netta mine. In drill cuttings, E bed is commonly indistinguishable unless the more crystalline type of limestone can be recognized in unaltered sections. In the cuttings examined from many drill holes, thicknesses of 5–50 feet within which the limestone is more coarsely crystalline have commonly been detected at some place in the interval 21–70 feet above the base of the Moccasin Bend. The wide range in position of this material, however, leaves some doubt as to whether it is everywhere stratigraphically equivalent. It lies lowest in the sections in the northeastern part of the mining field in Kansas, and highest in the northwestern part of the field lying west and northwest of Treece, Kans., and in the vicinity of the Garrett mine, 3 miles north of Treece. The regional variations in position are not regular, and there may be large differences in the drill holes on single 40-acre tracts. The occurrence of similar coarse-grained material at only slightly lower level in the G–H unit in the eastern part of the field in Oklahoma, as previously mentioned, casts further doubt as to stratigraphic continuity of a specific bed.

E bed is locally important because of its susceptibility to mineralization. Systematic exploration of this unit above old stopes has revealed new ore in some parts of the mining field in recent years.

Plate 4 also shows another fairly well marked unit near the top of the member. This is a thinly laminated zone of fine-grained limestone with alternating beds of chert, or of limestone with intermottled chert, and is tentatively correlated with C bed of Fowler and Lyden (Fowler, 1942). The thickness of the individual strata commonly ranges from a fraction of an inch to an inch, or in places, 2 inches. In the Crystal mine section, the thin-bedded material is interlayered with thicker beds of chert and limestone; the latter contains nodules and irregular masses of chert of varying sizes and shapes in different beds, from large melon-shaped nodules 8–12 inches thick and 2–3 feet long in some beds to small amoeboid masses in others. In some sections all the chert is light to pale gray, characteristic of the upper part of the Moccasin Bend; whereas in other sections, the Crystal mine section for example, much of the chert is blue gray, the color concentrated centrally or peripherally, or it is in small part uniform waxy gray. The supposed C bed unit has no distinctive lithologic feature that would make it definitely correlatable in drill cuttings. Perhaps much of the blue to pale chert that shows near the top of the Moccasin Bend in widely scattered drill cuttings is in this unit; but plate 4 shows that it is not persistent in the unit, and furthermore shows blue chert in places in both overlying and underlying units as well as in E and lower beds. Correlation is further complicated by the wide variation in thickness of the different units, C bed for example, being 4 feet thick in the Roanoke mine and 15 feet in the Crystal mine. The different beds making up the Moccasin Bend tend to show in different sections complementary thicknesses relative to other beds, so that overall thicknesses of the member approximate an average.

The B bed of Fowler and Lyden forms the top of the Moccasin Bend in the western part of the mining field but is absent in the Discard mine section at the eastern side of the field. Because its lithology is intermediate

between that of the underlying and overlying strata, its boundaries are in places hard to define, but the unit is believed to be present as far east as the Walker and Dobson mine tracts. It is a medium-to fine-grained, or locally rather coarse-grained, limestone, light gray or light brown, but more commonly pale to nearly white, and somewhat siliceous or cotton rocklike; the associated chert is generally light to pale gray, and cotton rocklike in places, but is in part light to pale blue in other places. Although the chert commonly averages 10–20 percent with fairly even distribution throughout the unit, it may be absent or nearly so, or it may make up a much larger percentage, with uneven distribution. It is nodular, lenticular, or in indefinite beds as much as 6 inches thick, with ill-defined or cotton-rock boundaries gradational into the limestone. B bed ranges in thickness from 0 in the eastern part of the mining field to 55 feet at one drill hole in the mineralized area southwest of Melrose, though 45 feet is the maximum in all other drill holes examined in this area. The maximum for the main mining field is 35 feet on parts of the See Sah and Blue Goose No. 2 tracts, but thicknesses approaching this maximum are scattered elsewhere in the western part of the field. The local variation in thickness is great, and there are many places even in the western part of the field where the unit is apparently absent or less than 10 feet thick. The minimum for the Melrose area is 10 feet in the single hole examined on the A. T. Wright tract.

In sections exposed along the Spring and Neosho Rivers, units above the G–H zone have not been recognized, although the general differences between G and H and higher beds persist. The upper part of the Moccasin Bend as formerly exposed in the Seneca graben block just above river level near the mouth of Spring Branch, 4 miles northwest of Grove, is unusual in that most of the limestone in the interval 17–40 feet below the top is coarse grained, or contains coarse grains in a fine-grained matrix. Much of it is fossiliferous. The basal 4 feet of this interval contains the coarse-grained material as layers or lenses in a fine-grained limestone. In the coarse limestones of the basal 12 feet are scattered crinoidal fragments of pink color, resembling those found in the overlying Chester limestones in some areas. The chert is subordinate, and of the usual pale colors, occurring irregularly marbled in the limestone or as small nodules or thin beds. The upper 17 feet of the Moccasin Bend at this locality is a fine-even-grained limestone, massively bedded, with little or no chert. Comparison with other sections along the Neosho in Delaware County suggests that virtually all the Moccasin Bend above the G–H zone is represented here. In a section formerly exposed on the river 2 miles to the southwest, the G–H zone is 23 feet thick. The next overlying strata here have a little glauconite which may be equivalent to that of E bed in the mining field.

Near Seneca, one of the cotton-rock beds 8–12 feet thick in the upper part of the Moccasin Bend forms the surface rock over rather extensive upland flats. This bed has been leached to yield the tripoli that has been extensively quarried in this vicinity. It contains masses of white chert that have to be discarded or avoided in working the deposits. In another tripoli-producing area about 3 miles southeast of Racine, the tripoli likewise is found on an upland flat. Thickness of the deposit here is only 5 feet.

Correlation and Age

The Moccasin Bend contains scattered fossils at different levels, and the occurrence of sponge spicules in the lower cherts has been mentioned. The unit, however, is rather poor in identifiable fossils, except for exposures of the upper part of the member at a few localities, some of which may represent the same or nearly the same bed. The notable exception is the uppermost bed present at the type locality. This bed is not exposed in place in the flattened slope at the top of the bluff, but the weathered boulders of mixed cotton rock and chert covering the outcrop in a vertical interval of 12 feet are large and angular, and obviously nearly in place. They contain well-preserved casts of numerous fossils. The zone represented is 59–71 feet above the base of the Moccasin Bend. What appears to be the same fossil zone is found also on and near the tops of hills elsewhere in the vicinity of Spring River in the northern part of the quadrangle. A similar fossiliferous zone in weathered cotton rock lies a few feet above the tripoli bed that is quarried near Seneca, Mo., and may be at the same horizon. The fossiliferous limestones formerly exposed in the Seneca graben near the mouth of Spring Branch cannot be related directly to the base of the member, which is below the river bed here, but comparison with adjacent sections suggests that they lie in the interval 25–60 feet above the base of the Moccasin Bend. Fossils are most abundant and best preserved in the top foot of this interval, which is probably the same fossiliferous zone as at the type locality.

Gordon comments (written commun., 1965) as follows on the fauna:

> Eleven fossil collections from the Mocassin Bend Member have yielded a fauna of sponges, corals, bryozoans, crinoids, blastoids, brachiopods, pelecypods, gastropods, and trilobites. They total about 70 species, the brachiopods predominant as usual. Long-ranging Osage brachiopods in the fauna include *Orthotetes keokuk* (Hall), **Avonia williamsiana* Girty, **Setigerites setiger* (Hall), **Echinoconchus biseriatus* (Hall), *Ovatia* aff. *O. pilei-*

formis (McChesney), **Rhynchopora palmeri* Girty, **Spirifer rostellatus* Hall, **S. tenuicostatus* Hall, **Pseudosyrinx keokuk* Weller, *Dimegalasma neglectum* (Hall), and *Eumetria verneuiliana* (Hall). Those species marked with an asterisk (*) are not known above this member. Typical early Meramec brachiopods in the fauna include *Spirifer bifurcatus* Hall, *Brachythyris subcardiiformis* (Hall), *Torynifer setiger* (Hall), *Cranaena sulcata* Weller.

The most abundant species in the fauna is *Setigerites setiger*, which reaches its greatest development in this member. *Brachythyris subcardiiformis* also is common, and the overlapping of its stratigraphic range with that of *S. setiger* is perhaps the most characteristic feature of this faunal zone. Several species appear to be restricted to this member, but most of them are rare. One of these, an undescribed species of *Heteralosia*, is fairly common.

Most of the fossils studied are from the richly fossiliferous zone in the upper part of the member. Of particular interest in the vicinity of Seneca are the crinoids which include, according to Kirk (oral commun., 1941), the genera *Actinocrinites, Linocrinus, Platycrinites, Dichocrinus,* and *Decadocrinus.*

The Mocassin Bend Member is believed to correlate approximately with the main part of the Warsaw Shale overlying the Geode beds at Warsaw, Ill.

Stratigraphic Relations

The Moccasin Bend Member overlies the Baxter Springs Member (J bed) conformably. It is overlain in at least a part of the mining field by the Quapaw Limestone of Warsaw or Salem age, with uncertain relations. Although the faunas do not indicate any appreciable hiatus, the gradual decrease in thickness of the Moccasin Bend from west to east suggests that there may be a diastemic break at the base of the Quapaw Limestone. However, the decrease in thickness is not all concentrated at this horizon; for lower units of the Moccasin Bend, the G–H zone for example, show thinning to the east and fanlike thickening to the west. A comparison of the sections exposed in the Roanoke mine, near Commerce, and the Discard mine, on the Oklahoma-Kansas line 1½ miles west of Spring River, indicates that, although the overall thickness of the Moccasin Bend is about 25 feet greater in the Roanoke mine, only 4½ feet of this increase is attributable to the presence of additional strata at the top of the Moccasin Bend section (pl. 4). Where the Moccasin Bend is definitely overlain by the Quapaw Limestone as identified by fossil collections, the increase in its thickness is from about 55 feet in the outcrop on Spring River southeast of Lincolnville to 80 feet in a shaft on the New Chicago No. 1 tract (SW¼NE¼ sec. 28, T. 29 N., R. 23 E.); but at what horizon this increase takes place is not known. Although even greater thickness to the west are probably beneath the Quapaw Limestone, this limestone is so similar to overlying Chester strata that the two were only imperfectly distinguished in the drill cuttings which furnished the only information on the subsurface strata west of the main mining field. If the Quapaw Limestone extends to the Melrose area, a possible diastemic break at its base could amount to the time equivalent of about 50 feet of limestone, the maximum thickness of the B bed of Fowler and Lyden.

At all places where overlying strata are preserved beyond the limits of the Quapaw Limestone, the Moccasin Bend Member is overlain unconformably by the Hindsville Limestone of Chester age, or the Krebs Group of Pennsylvanian age. Locally, the Moccasin Bend has been completely cut out beneath the unconformity at the base of the Hindsville (p. 51).

Quapaw Limestone

The only known outcrop of the Quapaw Limestone in the Wyandotte quadrangle is on the west side of Spring River at the bend southeast of Lincolnville, on the Pius Quapaw tract and the adjacent parts of tracts to the west and north. The locality lies 3 miles southeast of the town of Quapaw, for which the unit is herein named. Outcrops of the limestone appear in and bordering a shallow draw running back a quarter of a mile from the low river bluff. About 25 feet of the unit is present, consisting of medium to coarse-grained light-gray to brownish-gray crinoidal limestone, poorly stratified in beds 2–12 inches thick; the thicker beds of the unit predominate.

The outcrop was found and the faunal affinities were first recognized by James Steele Williams of the Geological Survey. The unit had previously been mapped by Siebenthal and Mesler (pl. 2) as part of the Hindsville Limestone. Because of lithologic similarity to the overlying Hindsville, the boundary between the two limestones has not been worked out in detail. Hence, the Quapaw Limestone may depart somewhat in thickness from that quoted, and it may crop out somewhat more extensively in the type area than shown on plate 2.

The Quapaw Limestone was also cut in a shaft sunk in 1937 on the New Chicago No. 1 tract (SW¼NE¼ sec. 28, T. 29 N., R. 23 E.), about 5 miles northwest of the outcrop. Here, 31 feet of the limestone containing the distinctive fauna rests on top of the Boone Formation and is overlain by the Pennsylvanian shales of the Krebs Group. The fauna is also represented in fossil collections made many years ago by George H. Girty from the dumps of shafts in the vicinity of Quapaw, about halfway between the Spring River outcrop and the New Chicago No. 1 shaft. The limestone is probably more widespread in the mining field than the known fos-

sil collections indicate, for it is lithologically so similar to parts of the overlying Hindsville Limestone that the two were not certainly distinguished where the only basis for judgment was the churn-drill cuttings. The presence of such a limestone unit unconformably below the Hindsville strata would explain otherwise anomalous variations from place to place in the thickness of the coarse-grained limestone just above the Boone. Drill holes in the Rialto structural basin 1 mile west of the New Chicago No. 1 shaft and nearly in line with the extension of the known occurrences of the Quapaw Limestone showed, below limestone of distinctive Hindsville types, a maximum of 72 feet of uniformly coarse-grained limestone that on the basis of lithology could all be assigned to the Quapaw Limestone. This thickness here may represent a structural distortion of the original stratigraphic thickness. Similarly, a drill hole in the structually lowest basin of the King Brand property (NE¼SW¼ sec. 11, T. 35 S., R. 23 E., Cherokee County, Kans.) showed, in the same geologic setting, 68 feet of limestone that might well be the Quapaw Limestone. In areas containing known Quapaw fossils, and in intervening areas, drill cuttings from this limestone contain at most only a trace of chert; but some of the local chert concentration in the limestone in other parts of the mining field could conceivably be in the Quapaw Limestone.

Correlation and Age

Gordon reports (written commun., 1965) as follows on the faunal content:

> Study of 18 collections of fossils from the Quapaw Limestone has resulted in identification of about 55 species including bryozoans, brachiopods, pelecypods, gastropods, and trilobites, with the brachiopods predominant; the unit also contains crinoid, echinoid, and fish remains. Of those brachiopods that appeared in the Boone Formation at the beginning of Keokuk time, that is, in the Grand Falls Member, only *Orthotetes keokuk* (Hall), *Echinoconchus biseriatus* (Hall), *Ovatia* aff. *O. pileiformis* (McChesney), *Dimegalasma neglectum* (Hall) and *Eumetria verneuiliana* (Hall) remain. The rest are all species not known below the base of the Warsaw, that is, below the Short Creek Oolite Member.
>
> The Warsaw fauna includes *Perditocardinia* n. sp., *Orthotetes kaskaskiensis* (McChesney), **Platyselma echinatum* Gordon, **P. quapawense* Gordon, *Crossacanthia perlamellosa* Gordon, *Labriproductus wyandottensis* Muir-Wood and Gordon, *Allorhynchus acutiplicatum* Weller, *Spirifer bifurcatus* Hall, *S. washingtonensis* Weller, var., **S. lateralis* Hall, **S. incertiformis* Girty, var., *Brachythyris subcardiiformis* (Hall), *Torynifer setiger* (Hall), *Eumetria lata* Snyder, **Cleiothyridina densa* (Hall), *Composita globosa* Weller, *C. trinuclea* (Hall), and **Beecheria obovata* (Hall and Clarke). Species that are restricted to the Quapaw Limestone in the Wyandotte quadrangle are marked with an asterisk (*).
>
> The restriction of *Spirifer lateralis* to this formation is puzzling, because elsewhere, this typical species ranges much lower in the Warsaw. In the Joplin district in nearby Missouri, this species appears to be present in the Baxter Springs Member. Its absence in this member in the Wyandotte quadrangle may be connected with development of a silty facies in the Baxter Springs Member in parts of this quadrangle, which brought conditions that were unfavorable for the development of *Spirifer lateralis*.
>
> Small spiny productoid brachiopods are a rather distinctive feature of this fauna, and the relatively common species *Platyselma echinatum* Gordon may be considered as the characteristic species of this faunal zone.
>
> The Quapaw fauna contains elements that one might consider late Warsaw in age, but the possibility of a Salem age for this formation cannot be eliminated at present. The well-know Salem fauna that occurs in the oolite in southern Indiana is a facies fauna. Where the oolitic facies is not present, the fauna of the Salem Limestone is much like that of the upper part of the Warsaw. Several species that are used to differentiate the faunas of these two formations are not in the Quapaw Limestone, so for the present it is best to consider the Quapaw as late Warsaw or Salem in age.

Stratigraphic Relations

The Quapaw Limestone is probably separated from the underlying Boone formation by a diastemic break, as discussed in the account of the Moccasin Bend Member. At the known outcrop the Quapaw is overlain by the Hindsville Limestone with presumed unconformity, though the contact has not been studied. At the New Chicago No. 1 shaft it is overlain unconformably by the Pennsylvanian shale.

CHESTER SERIES

The Chester Series is represented in the Wyandotte quadrangle by three formations, comprising a maximum stratigraphic thickness of about 200 feet, although this thickness is nowhere present in any one section. The basal formation, the Hindsville Limestone, is a limestone with minor sandstone lenses. It grades through an indefinite boundary into the middle unit, the Batesville Sandstone, which is a marine sandstone with abundant interbedded limestone and some shale. The upper formation, the Fayetteville Shale, is a marine shale with some interbedded limestone. All are fossiliferous, especially the upper unit. The series is unconformable on the Boone Formation or on the Quapaw Limestone where present, and, in turn, is overlain unconformably by Pennsylvanian strata. In places, the whole series was removed by erosion before the deposition of the Pennsylvanian, in other places, near-maximum thicknesses of Chester strata are preserved in sharply defined structural basins and solution slumps that formed in the interval between the deposition of the Chester and the beginning of Pennsylvanian deposition.

Hindsville Limestone

The Hindsville Limestone is the basal, more limy facies of an interfingering sandstone-limestone sequence in which the sandstone first becomes conspicuous at some distance above the base. As the sandstone is highly lenticular, the horizon at which it becomes prominent varies from place to place. Thin beds of sandstone or sandy limestone may appear in some sections near or at the base of the Chester Series, whereas closely adjacent sections have limestone in equivalent parts of the section. Compared to Chester strata farther to the east on the south side of the Ozark dome, the facies in the Wyandotte quadrangle is predominantly limestone; and had not the sequence been studied earlier in the eastern area where the limestone forms a minor basal member of a predominantly sandstone interval, the Batesville Sandstone, it is doubtful that two stratigraphic units would have been recognized. Indeed, Snider (1915) and Weidman (1932) do not subdivide the sequence stratigraphically, classing it all as Mayes formation. However, Siebenthal and Mesler (pl. 2) have mapped the upper sandstone-bearing facies separately from the limestone below, and this classification is here followed, though in many places the distinction is rather artificial. In any given section, the contact is drawn at the base of the lowest prominent sandstone bed, which is usually at least 4 feet thick. Although the stratigraphic horizon is not everywhere the same, details of intertonguing cannot be traced on the surface, nor could they be shown at the scale of the mapping.

The Hindsville is the surface rock over rather extensive areas in the western half of the Wyandotte quadrangle, and in addition it has a limited distribution in the eastern half. In a setting of low northwesterly regional dip and rolling prairie terrain, its main outcrop shows a pattern that is rather intricate in detail within a broad northeast-southwest belt lying west of Spring River and west of the Neosho below its junction with Spring River. The belt extends from Lincolnville on the northeast through Afton on the southwest and includes both Miami and Fairland. Within this belt the Hindsville outcrop fingers irregularly to the northwest up the shallow stream valleys, and more broadly to the southeast on the flat upland divides, with outliers beyond the band of continuous outcrop. Northwest of the main outcrop belt the limestone is exposed in inliers along Tar Creek and its east tributaries, west and southwest of Quapaw. Branson (Reed and others, 1955, pl. 1) maps a small inlier of Hindsville Limestone in the bed of the Neosho River, 6 miles northwest of Miami. This outcrop must be visible only at the very lowest water stage of the river, for at other stages it is indicated only by a riffle in the water surface. That Chester strata are indeed close to outcrop here is shown by the records of drilling that reached the Precambrian granite on the adjacent Demo tract. The west hole (pl. 2) reached "dense gray limestone" below the terrace alluvium somewhere in the interval 12–20 feet below the surface; the east hole intersected "limestone boulders" at a depth of 10 feet, and "medium-grained lime" at 15 feet. The buried granite hill at shallow depth may well be the cause of the doming which accounts for these strata near the surface, for it could act as a competent block in transmitting any structural stresses, and the warps might well come close to its boundaries.

Southeast of Afton the Horse Creek anticline, which is asymmetric, with the steeper dip on the southeast flank, carries the Hindsville and overlying strata down on this flank into a flat shallow syncline in which the Hindsville outcrop rings and defines the synclinal basin. The Hindsville is preserved in the same structural setting in relation to the extension of the Horse Creek anticline east of the Neosho and north of the Elk River, but here the synclinal basin is shallower; hence little is preserved above the Hindsville.

Other outcrops east of the Spring and Neosho Rivers are chiefly outliers on the broad flat drainage divides. The most extensive of these outliers are on Cowskin Prairie, between Grove and South West City. In such outliers, both east and west of the rivers, the Hindsville may be the only formation involved, or overlying formations may cap the flat uplands. Smaller patches of the limestone are also preserved east of the river in large sinkholes developed in limestones of the underlying Boone. The Hindsville also occurs in places within the narrow Seneca graben block.

The Hindsville is present west of the main belt of outcrop in some of the mines where it has been mineralized. Prospect drilling has also revealed it in many of the holes drilled in the mining field, though in some areas it is missing either as a result of pre-Pennsylvanian erosion or of post-Pennsylvanian underground solution.

Thickness

As the top of the Hindsville is an indefinite stratigraphic boundary, the thickness varies from place to place. On Horse Creek 1½ miles west of Needmore, the thickness formerly exposed was about 15 feet. On the left bank of the Neosho 2 miles east-northeast of Needmore, the thickness is 13 feet; and several former exposures on the right side of the river for 2 miles below Sailboat Bridge had thicknesses of 15–30 feet. About 40 feet of Hindsville is present at the top of the bluff on the east side of Spring River 3 miles southeast

of Moccasin Bend, and about 57 feet in a section 2½ miles northeast of here, on the south side of Flint Branch. The unit is 35–45 feet thick near Miami, as measured at several places between the Kansas, Oklahoma & Gulf Railway bridge and the banks of the Neosho near the mouth of Tar Creek.

In the mining field the thickness is only partly known because lithologic distinction from the underlying Quapaw Limestone was not made in most of the drill cuttings that were available for study. In some areas where a little sandstone or other lithologic features characteristic of Hindsville strata occur at the base of the coarser grained limestone sequence above the Boone, the presumption seems reasonable that only Hindsville strata are represented. Deductions as to stratigraphic thickness must be based on several drill holes, for stratigraphic cutouts due to structural squeezing can be demonstrated under favorable conditions in the limestone strata of the area, and cutouts or condensations in thickness due to undetected solution are omnipresent possibilities. In addition to the original variations in stratigraphic thickness of the Hindsville, variation has been produced by the erosional unconformity at the base of the Pennsylvanian strata, which has in places truncated below the base of the Hindsville Limestone.

On the John Beaver tract the Hindsville has for the most part been extensively truncated by the pre-Pennsylvanian unconformity, but a few drill holes along the syncline (Miami trough) traversing the tract indicate an original thickness of about 45 feet. Several drill holes on the Blue Goose No. 2 tract indicate that the Hindsville Limestone is here 25–40 feet thick. Several drill holes in unaltered ground on the Maxine tract within an area 350 feet across show limestone thicknesses of 30–55 feet between the Boone Formation and Batesville Sandstone; this gives maximum posible thicknesses for the Hindsville, which may be diminished by any undetermined thickness of Quapaw Limestone that may be present.

On parts of the Scammon Hill tract the Batesville is either directly on top of the Boone or only 3–10 feet of Hindsville Limestone intervenes. As the underlying Moccasin Bend Member of the Boone has been leached of all limestone, it appears that all or a large part of the Hindsville Limestone was here dissolved out at the same time. There are many localities where intense leaching by ground waters has removed all the Hindsville from beneath the Pennsylvanian shaly strata.

The thickest interval of Hindsville plus probable Quapaw Limestone, penetrated in drill holes whose cuttings were examined in our fieldwork, was at the east end of the Rialto structural basin, on the Barbara J. tract. Here, one hole penetrated 147 feet of limestone between the Boone Formation and Batesville Sandstone. In view of the location of the basin a mile beyond but nearly in line with the proven occurrences of the Quapaw Limestone, the presumed presence of this limestone below the Hindsville in the basin is a reasonable assumption to account for the excessive thickness of limestone present here. Only traces of sandstone were found in the upper part of the limestone in an interval 30–40 feet below its top. However, the Hindsville part of the section is more than 40 feet thick, for several neighboring drill holes showed thin beds of sandstone or other indications of Hindsville lithology over an interval of 60 to at least 85 feet at the top of the limestone. Possibly thickening of the limestone in the sharply defined structural basin may have distorted stratigraphic intervals or disturbed the Batesville contact in this area, for contrasts between adjacent holes are particularly great.

The full thickness of the Hindsville is preserved in the mining field only in those structural basins that antedated truncation by the erosional unconformity at the base of the Pennsylvanian. This unconformity cut below the top of the Hindsville in most of the mining field, but details as to the thickness remaining are not known because of the Quapaw Limestone complication. The Hindsville has been completely eroded in some areas of structurally high ground on the west side of the Miami trough and Bendelari monocline as, for example, on parts of the Robinson, Stanley, Pelican, and Gordon No. 3 tracts, where the Pennsylvanian strata rest on the Boone. The same situation exists in a small area just north of the State line in the Melrose area, in W½ fractional sec. 15. Areas from which the Hindsville has been completely eroded over Quapaw Limestone may be extensive in parts of the mining field, but the only locality where this is definitely demonstrated is at the shaft on the New Chciago No. 1 tract which yielded Quapaw fossils from all the limestone interval between Boone and Pennsylvania strata.

Character

The limestone in the Hindsville is dominantly medium to rather coarse grained and crinoidal, but contains some fine-grained to dense beds. It is generally light gray or light brown, though darker in some sections and paler in others. Some of the crinoidal grains in local areas are pink, flesh colored, or yellow. Some of the dense limestone is pyritic and weathers buff. In places the limestone may be slightly glauconitic. Oolite is common, but exhibits nowhere near the perfection found in the Short Creek Oolite Member of the Boone. The oolitic spherules are commonly in a nonoolitic matrix, and a larger

percentage than in the Short Creek are oblong rather than round. The oolite occurs haphazardly in the limestone, has no persistence along the bedding, and is not recognizably massed into definite beds. The beds of the limestone are usually 2 inches to 2 feet thick and commonly weather slabby on the outcrop, but in places may be thin and platy. Locally, the crinoidal or oolitic limestone is crossbedded.

In some places in the mining field the Hindsville contains appreciable chert, unevenly distributed in all parts of the section but commonly concentrated at the top where capped by the Pennsylvanian. In contrast to that in the Moccasin Bend Member of the Boone, the Hindsville chert is usually grayer, more translucent, and commonly has more organic textures. It may have the granular texture of replaced limestone. Some of the chert, however, is indistinguishable from that of the upper part of the Boone, and there may be all gradations between the two types. The chert colors may include some translucent blue, pale blue, or light brown, in addition to the dominant gray. Where seen at the top of a 9-foot wedge of Hindsville and just below the Pennsylvanian shale in the Crystal mine, the chert is in large nodules, as much as 1½ feet long and several inches thick.

The occurrence of sandstone or sandy limestone streaks in the Hindsville is widespread, though present only in some sections, and generally in thin beds and in small amount. The contained sand is fine grained, commonly pyritic, and light greenish gray where fresh, but weathers buff or brown. It is identical with that in the overlying Batesville Sandstone. The sandstone is lenticular, but locally a bed even near the base of the Hindsville may become as thick and conspicuous as individual beds in the overlying Batesville. The sandstone is usually thin bedded, locally ripple marked, and both it and the sandy limestone are commonly crossbedded. The Hindsville also contains a little shale at scattered localities, in thin seams between limestone, but locally as much as 7 feet thick. This shale is green, bluish green, or gray, commonly pyritic, in part sandy, and weathers to drab or buff. It may also occur as small pellets in the limestone.

A chert-pebble conglomerate in limestone is described by Weidman (1932, p. 18) as occurring at the base of the Hindsville near the Kansas, Oklahoma & Gulf Railway bridge at Miami. The pebbles are irregular in shape but usually have smooth surfaces, and range in size from less than an inch to 2 or 3 inches in diameter. They are interpreted by Weidman as having been derived from weathering and erosion of the underlying Boone during the time interval represented by the unconformity at the base of the Chester Series. A similar conglomerate was formerly exposed in Delaware County 2 miles southwest of Sailboat Bridge, in a quarry from which rock for the Grand River dam was obtained. The basal 1–1½ feet of the Hindsville is here a coarse-grained limestone which contains chert pebbles, the type of chert resembling that found in the upper part of the Boone. The pebbles are subangular to dominantly well rounded, averaging ½–1 inch in diameter, with a maximum of about 2 inches. Although this conglomerate is apparently conformable on underlying beds in the quarry, at 1 mile to the southeast the base of the Chester is undulatory, with a relief of 1 foot in 3, and without the chert pebbles. The features described at these two localities are further evidence of an erosion unconformity at the base of the Chester.

The Hindsville is characterized by great lithologic variability from place to place, so that it is impossible to trace individual beds or zones for any distance.

Correlation and Age

The Hindsville Limestone was named as the basal member of the Batesville Sandstone by Purdue and Miser (1916) from a locality in the Eureka Springs quadrangle, Arkansas. It was mapped by them into the northwest corner of that quadrangle, which lies 28 miles from the southeast corner of the Wyandotte quadrangle. Although the Batesville Sandstone had been earlier mapped in the Fayetteville folio which occupies the existing gap, the limestone at its base had not then been differentiated from the Boone. However, the stratigraphic position of this limestone, its relation to overlying and underlying formations, its thickness, general lithologic character, and the occurrence of a basal conglomerate are so similar in the Wyandotte and Eureka Springs quadrangles as to leave little doubt of stratigraphic identity.

On the faunal content and its relation to that of other stratigraphic units, Gordon reports (written commun., 1965) as follows:

> The fauna of the Hindsville Limestone is large. About 200 species and subspecies of invertebrate fossils have been recognized in 43 collections from the Wyandotte quadrangle, including mollusks, brachiopods, bryozoans, corals, echinoderms, and trilobites, in descending order of abundance of species. In actual numbers of individuals and species at most localities, the brachiopods are the most abundant forms, but mollusks are found in all beds; locally, in lenses of particularly pure limestone the number of small mollusks is much greater and the brachiopods less.
>
> Common brachiopods characteristic of the Hindsville Limestone are *Orbiculoidea caneyana* (Girty), *Perditocardinia* aff. *P. dubia* (Hall), *Orthotetes subglobosus batesvillensis* Girty, *Chonetes tumescens* (Easton), *Diaphragmus cestriensis* (Worthen), var., *Inflatia inflata* (McChesney), *Flexaria arkansana*

(Girty), *F. multilirata* (Girty), *Ovatia elongata* Muir-Wood and Cooper, *"Productus" siebenthali* Girty, *Camarotoechia purducei* Girty, *Spirifer* aff. *S. pellaensis* Weller, *Punctospirifer transversus* (McChesney), *Reticulariina subspinosa* Weller, *Syringothyris* aff. *S. aequalis* Easton, *Torynifer setiger* (Hall), *Eumetria vera* (Hall), *E. verneuilana* (Hall), *Cleiothyridina elegans* Girty, *C. sublamellosa* (Hall), and *Composita subquadrata* (Hall). Common and characteristic mollusks include *Sphenotus washingtonensis* Girty, *Leptodesma* aff. *L. nasutum* (Herrick), *Myalina elongata* Girty, *M. illinoisensis* (Worthen), *Schizodus depressus abruptus* Girty, *Sulcatopinna arkansana* (Weller), *Wilkingia walkeri* (Weller), *Straparollus (Euomphalus) planidorsatus* Meek and Worthen, and *Orthonychia* cf. *O. acutirostris* (Hall). The Hindsville molluscan fauna aggregates about 45 species of pelecypods and 50 species of gastropods, many of them as yet undescribed. The trilobite *Paladin mucronatus* (Girty) also is fairly common.

The only species known to carry over from the rocks of Warsaw and possible Salem age beneath the unconformity into the Hindsville are the brachiopods *Torynifer setiger* (Hall), *Eumetria verneuilana* (Hall), and *Cleiothyridina sublamellosa* (Hall).

The Hindsville fauna is about the same as that of the Hindsville Member of the Batesville Sandstone in the region of its type locality in nothern Arkansas. It is early Chester in age. South of the Wyandotte quadrangle, from the vicinity of Spavinaw, Okla., southward, the Hindsville Limestone rests on the various members of the Moorefield Formation (Huffman, 1958, p. 63) of Meramec age. In the Wyandotte quadrangle the Moorefield is represented, at least in part, by the unconformity beneath the Hindsville Limestone.

Five collections from the Wyandotte quadrangle contain brachiopods, such as *Auloprotonia aulacophora* Muir-Wood and Cooper, *Camarotoechia purduei agrestis* Girty, *Leiorhynchus carboniferum* Girty, *Moorefieldella* aff. *M. eurekensis* (Walcott), *Spirifer arkansanus* Girty, and *Spirifer martiniiformis* Girty, generally considered to be typical of Moorefield rocks and not found elsewhere in the Hindsville. Absent in the same collections are ubiquitous Chester species such as *Diaphragmus cestriensis* (Worthen) var. and *Reticulariina subspinosa* (Weller). However, some characteristic Hindsville brachiopods, such as *"Productus" siebenthali* Girty and *Syringothyris* aff. *aequalis* Easton, do occur. Also present are *Leiorhynchus carboniferum* Girty, and *Moorefieldella* aff. *M. eurekensis* (Walcott) which are known in the Fayetteville Shale. Two alternative interpretations of the faunal evidence are possible: either (1) these faunules occur in calcareous outliers of Moorefield age to the north of the known outcrop area of the Moorefield formation in Oklahoma and are overlapped by the Hindsville Limestone or (2) the faunules are recurrent in rocks of Hindsville age.

In the two areas in the Wylandotte quadrangle where the Moorefield faunules were obtained—one along the Neosho River northwest of Grove near the mouth of Spring Branch, and now submerged beneath the waters of Lake O' the Cherokees and the other near Lincolnville—the fossils were obtained from the upper part of the Hindsville, just below the Batesville Sandstone. It is difficult, therefore, to regard these occurrences of fossils generally regarded as Moorefield forms as being anything but extensions of their stratigraphic ranges into rocks of Hindsville age.

The Hindsville is the lower part of the Mayes Formation as this term was used in Ottawa County by Weidman (1932). Snider (1915), who introduced the term "Mayes," mapped it over a much larger area and realized that the lower part of it south of the middle of the Pryor quadrangle, Oklahoma, was Moorefield, and thus older than the part preserved in the Wyandotte quadrangle where his usage was followed by Weidman. According to Huffman (1951, p. 6), the Hindsville of Brant [4] (and presumably also Brant and Fitts, 1941) in Mayes County, Okla., is a member of the Moorefield and thus below the true Hindsville; and Brant's term "Grand River" is used to designate the true Hindsville plus its overlying Batesville equivalent (see also Slocum, 1955, p. 12). The Carterville Formation, which occurs only in sinkholes in the Joplin area (Smith and Siebenthal, 1907, p. 5–6), is the equivalent of the Hindsville, Batesville, and Fayetteville, though probably only one or two of these are represented in any given sinkhole.

Stratigraphic Relations

At one locality southeast of Quapaw, the Hindsville overlies the Quapaw Limestone of late Warsaw or Salem age; and although the contact has not been studied, it is undoubtedly one of unconformity. Elsewhere in the Wyandotte quadrangle the Hindsville overlies the Boone Formation with an erosion unconformity revealed at several places. Although generally the erosion surface lies on the upper part of the Moccasin Bend Member of the Boone, at at least one locality near the Ottawa-Delaware County line nearly 2 miles east of Turkey Ford, all the Moccasin Bend and Baxter Springs Members have locally been truncated. Here, at a quarter of a mile to half a mile up the small branch from the county-line road crossing, 7½ feet of fine sandy thin-bedded limestone lies directly on the Short Creek Oolite Member, and is overlain by 6 feet or more of fine-grained light-gray to drab argillaceous limestone which weathers slabby. Sparse fossils collected by Mackenzie Gordon, Jr. (written commun., 1965), from this limestone contained the following:

Ovatia elongata Muir-Wood and cooper, *Echinoconchus* n. sp. A, and *Spirifer* n. sp. A. Careful preparation and examination of the *Ovatia* showed that the spine pattern along its hinge is that of the common Chester species and not that of the common Boone species. The *Echinoconchus* and the *Spirifer* also agree more closely with the Chester forms of these genera than with other species of the same genera from the Boone Formation. It can be said with reasonable assurance, therefore, that the rocks overlying the Short Creek Oolite Member of the Boone Formation in the section belong to the Hindsville Limestone of early Chester age.

At other exposures in this general vicinity, thin remnants of the Baxter Springs Member are preserved

[4] R. A. Brant, 1941, Stratigraphy of the Meramec and Chester series of Mayes County, Oklahoma: Tulsa Univ., unpub. M.S. thesis.

below the Hindsville. Thus, on the west side of the branch at the county line-road crossing, 1–2 feet of questionably L bed chert and 6 inches of K bed, glauconitic limestone with small phosphate pebbles, intervene between the Short Creek Oolite Member and the Hindsville Limestone, which here has lost the sandy component from its basal beds.

The maximum time interval represented by the unconformity at the base of the Hindsville thus embraces virtually all the Warsaw (= basal Meramec) and the rest of the Meramec Series.

Subsurface leaching of limestone strata in the Boone to form underground caves, which process persisted through several recurrent intervals of the geologic history, was probably initiated during the erosion interval represented in the pre-Hindsville unconformity. Drilling in the mining field has revealed several places where Chester strata occur well down in the Boone, below a thickness of Boone strata penetrated by the drill at higher levels. As long as the only evidence is obtained from drill cuttings, it is not obvious whether such Chester material was initially deposited in a cave in the Boone, or has slumped, after solidification, somewhat laterally into such a cave formed at a later period. However, in the Mary Jane mine northeast of Quapaw, a cave filling 100 feet across and as much as 10 feet thick in the upper part of K bed is exposed in the east wall of the mine near the north shaft. The filling material consists dominantly of hard green to greenish-gray clay shale, finely laminated, containing gradational layers of fine- to medium-grained gray limestone as much as 5 inches thick, some limy siltstone, and a bed of sandstone ranging in thickness from a thin wedge to 1 foot. Chert fragments, some more than 1 inch long, lie parallel to the bedding in the thickest limestone seam. The shale contains numerous thin crystalline blades of marcasite and smaller granules of poorly crystalline pyrite. From the undeformed character of the shale laminae and the tightness with which the shale fits, with undisturbed contact, against the limestone cave walls, there is no question as to the existence of the cave before the shale was emplaced. As the shale contains no fossils, however, its stratigraphic affinities have to be surmised. It is very much like the shales of the Batesville and Hindsville, and totally unlike the black shales that characterize the Pennsylvanian rocks. Furthermore, the amount of interbedded limestone indicates a limy environment in the source sediments, which is, again, like the Chester and unlike the Pennsylvanian rocks. Unless a special source is postulated that is not represented in the sedimentary record, which seems unlikely, the most probable source is from the materials deposited during the early stages of Chester deposition, some of which are indistinguishable from this cave deposit. As the base level of erosion in the post-Boone pre-Chester interval had nowhere in the mining field cut, so far as known, below the nearly horizontal plane that marked the top of the Boone at the beginning of the Chester sedimentation, it is probable that this cave formed below the ground-water table.

In places, particularly on the uplands east of the Spring and Neosho Rivers, the Hindsville fills large sinkholes in the upper part of the Boone. These sinkholes are believed to have formed well after the deposition of the Hindsville. They give a solution slump contact of Hindsville against some horizon below the stratigraphic top of the Boone, and probably in most places the basal part of the Hindsville has also been dissolved. The stratigraphic relationships involved in such contacts are probably in all places obvious.

The Hindsville in most places is overlain conformably, and with transgressing interfingering contact, by the Batesville Sandstone. The transition comes at an indefinite level above the base of the Hindsville where the sandstones first become prominent. As these sandstones are lenticular, the stratigraphic level of the transition varies from place to place, in general occurring progressively higher in the section between the southern part of the Wyandotte quadrangle and the mining field. In parts of the outcrop stretch from Narcissa to and beyond Afton, the sandstones fringe out and are replaced laterally by limestone. The upper part of the Hindsville in such places is the lateral equivalent of the Batesville Sandstone and may be overlain conformably by the Fayetteville Shale.

In the subsurface over a large part of the mining field and also on the outcrop near Miami, the Hindsville is overlain unconformably by the Krebs Group of Pennsylvanian age.

Batesville Sandstone

The Batesville Sandstone is the upward continuation of the Hindsville Limestone in which sandstone becomes generally prominent. However, the Wyandotte quadrangle appears to lie near the western fringe of the sandstone; hence in those sections that are complete, or nearly so, the sandstone is subordinate to limestone and shale. The sequence might well have been termed "Batesville Formation" instead of Batesville Sandstone; but as the fringe area involved is narrow in comparison to the wide extent of Batesville to the east in Arkansas, and as it is doubtful that equivalent strata extend very far west beneath the Pennsylvanian cover, it seems best to carry the terminology that prevails

farther east at least through this western limit of outcrops surrounding the Ozark dome.

The limits of the Batesville Sandstone outcrops are practically coextensive with those previously given for the Hindsville. On many of the broad flat divides, the more competent sandstones of the formation cap extensive benches. On others, however, the Batesville tends to be reduced by erosion to isolated outliers capping the low hills in which only the lower part of the formation is preserved. In parts of the outcrop strip of Chester strata between Narcissa and the quadrangle boundary southwest of Afton, the Batesville is missing from its normal position between the Hindsville and Fayetteville. It is presumed to be represented here by limestones of equivalent age that have been mapped with the Hindsville. The occurrence of Batesville in sinkholes on the flat uplands east of Spring and Neosho Rivers is more widespread than that of the Hindsville, particularly in the northeast corner of the Wyandotte quadrangle, for the sandstone that characterizes the Batesville is more likely to be preserved where the underlying limestone has been dissolved out. Hindsville and Batesville are inextricably jumbled together along parts of the Seneca graben.

In the western part of the mining field, only the lower part of the Batesville is preserved in certain structural basins that antedated the pre-Pennsylvanian erosion, but eastward in the field, equivalent strata are also preserved in small areas outside such basins. Not enough drill cuttings have beeen examined by us to define in detail the extent of the Batesville in the subsurface. Enough is known of the distribution, however, to indicate that the Batesville was eroded in pre-Pennsylvanian time from a much larger area of the mining field than that in which it is now preserved.

Thickness

At no place on the outcrop is a complete and satisfactory thickness of the Batesville exposed, although former poor exposures along the Seneca garben on the west side of the Neosho about 2 miles southwest of Sailboat Bridge suggest a thickness of about 40 feet. Harris (1956, p. 263) records 38½ feet at a locality 3 miles southeast of Miami. At most other places, only the basal few feet of sandstone, and perhaps some interbedded shales or limestones, are exposed, and the upper part of the unit is eroded. In such erosion remnants, the remaining thickness of Batesville rarely exceeds 15 feet.

The thickness revealed by drilling in the mining field ranges from 0 to 70 feet. As the top of the unit was nearly everywhere eroded beneath the unconformity at the base of the Pennsylvanian, the true original thickness of the Batesville is not known. West of the Bendelari monocline and Miami trough, it was completely removed except in a few small structural basins. In one such basin on the Federal Jarrett, 60 feet of strata penetrated by a drill hole and referred to the Batesville may not be an accurate stratigraphic thickness because of structural complications, but this figure at least is an approximation.

On the John Beaver tract, 15 feet of Batesville was cut in certain drill holes on the east flank of the Miami trough, though no Batesville was cut in several drill holes more nearly on the present trough axis. A similar condition prevails on the Ritz tract where the thickest Batesville found (45 ft) lies on the east flank of the trough. The Batesville is as much as 70 feet thick along the trough on the Scammon Hill tract.

Thicknesses of as much as 45 feet occur in relatively flat-lying strata on the Blue Goose No. 2 tract, well east of the Miami trough. The thickest occurences are in relatively insignificant structural basins or high on the flank of the larger structural basin that is centered on the South Side tract. Possibly thicker remnants are preserved on the South Side, but no drill cuttings from this tract were seen. At least 60 feet of Batesville is preserved in the Rialto basin (Barbara J. tract), and 65 feet is preserved in the south arm of this basin (northeast corner of Admiralty tract). The thickest Batesville found on the Barr tract, about 50 feet, is on the east flank of a pre-Pennsylvanian basin at the north center of the tract.

Local maxima in basins in the eastern part of the field are 40 feet on the Crane and 30 feet on the Thomas tract (just south of Euterpe mine). A maximum of 22 feet on the Betsy Greenback, however, is in a small area of nearly flat structure. This occurrence departs from the more common one in basins.

The Batesville is absent over most of the area of tracts for which thicknesses have been cited, and it is preserved on these tracts only in local spots where some special condition prevails.

Character

The Batesville Sandstone in the Wyandotte quadrangle is like the Hindsville limestone except that it is somewhat sandier, and it also contains a relatively higher percentage of oolite. The types of lithology are the same in the two formations and only the proportions differ. The different lithologic types in the Batesville are interbedded, and in the most complete sections limestone predominates over the sandstone and shale.

The sandstones of the Batesville are lenticular, and commonly grade to limestone both along and across the

bedding. They occur irregularly in the limestone in beds a few inches to as much as 32 feet thick, though they generally are less than 15 feet. In many sections only one or two sandstone beds are present. The thicker beds do not continue very far laterally before they break up into thinner lenses. For example, on the Blue Goose No. 2 tract one drill hole shows a 32-foot sandstone bed that has broken into four beds separated by limestone in another drill hole less than 300 feet away. Commonly, sandstones cannot be correlated between drill holes with spacing as close as 100–200 feet. The base of the Batesville, at the lowest significant sandstone, varies in stratigraphic position from place to place, as indicated in the discussion of the Hindsville thickness. Nevertheless, in some areas the sandstone tends to be concentrated and beds tend to be thickest in a fairly definite stratigraphic zone whose base arbitrarily becomes the base of the Batesville. In the section along the Neosho River 2 miles southwest of Sailboat Bridge, in Delaware County, the basal 10 feet is dominantly sandstone, and the remainder of the Batesville, 30 feet, is limestone.

The Batesville sandstones are medium to fine grained and white, light gray or pale green where fresh, but weather buff, brown, and reddish. They may contain a little gray chert locally, and have been converted to gray quartzite in mineralized ground. The sandstone grains include a variable amount of feldspar, locally as much as 26 percent, according to Weidman (1932, p. 20). The sandstones are commonly thin bedded, or include massive beds as much as 3 feet thick, and they may show ripple bedding or low-angle crossbedding. The more vigorous wave action at the floor of the Chester sea that led to the wide spreading of sand at the beginning of Batesville time probably also accounted for the increased amount of oolite in the calcareous sediments. In outcrops formerly exposed on the southeast side of the Horse Creek anticline southeast of Afton and northwest of Grove, a sandstone near or at the base of the Batesville so intergrades with oolitic limestone that in many sections a recognizable oolitic sandstone could be used as a horizon marker.

In the coarser limestone of the Batesville, pink, flesh-colored, and yellow crinoidal stem fragments, locally as much as three-quarters of an inch in diameter, occur more commonly than in the Hindsville. This color feature seems to be related in some way to the greater abundance of oolite. Chert is negligible in the limestones of the Batesville interval.

Green shale, grading in places to gray, is commonly interbedded with, or forms thin partings in, the sandstone, and the two commonly intergrade. The shale also shows gradation to both the fine- and coarse-grained crinoidal limestone. Some of the shale is pyritic.

Correlation and Age

The limestone and limy sandstone of the Batesville contain numerous marine invertebrate fossils and also scattered fish teeth. On the basis of this fauna, the unit is correlated with the Batesville of northern Arkansas which is of Chester age (Purdue and Miser, 1916). Mapping is nearly continuous from the Wyandotte quadrangle through the Fayetteville quadrangle, Arkansas (Adams and Ulrich, 1905), to the Eureka Springs and Harrison quadrangles, Arkansas, where the Batesville has been particularly well studied (Purdue and Miser, 1916).

Gordon (written commun., 1965) has supplied the following comments on the fauna:

> The fauna of the Batesville Sandstone in the Wyandotte quadrangle, as determined from 23 fossil collections, aggregates not quite 90 species and subspecies. These include mollusks, brachiopods, bryozoans, corals, trilobites, and echinoderms in decending order of number of species. The fauna is much like that of the Hindsville Limestone, except that the species with Moorefield affinities present in the Hindsville are not represented in the Batesville, nor are several other Hindsville species, such as *"Productus" siebenthali* Girty and *Syringothyris* aff. *S. aequalis* Easton. Also, the Batesville Sandstone lacks the limestone lenses crowded with small mollusks that so greatly increase the number of species in the Hindsville fauna. Aside from these differences, the Batesville fauna includes most of those listed on pages 59–60 as typical of the Hindsville.

Stratigraphic Relations

The Batesville Sandstone overlies the Hindsville Limestone conformably with transgressive boundary and intertonguing of beds. Locally, where the definitive sandstones were not deposited, an upward extension of the Hindsville limestone is the lateral equivalent of the Batesville.

Where complete underground leaching of the Hindsville has occurred in post-Chester time, the Batesville may show a false stratigraphic contact with the underlying Boone, as on parts of the Scammon Hill tract in the mining district, and in many of the small outlying occurrences, partly in sinkholes, east of the Spring and Neosho Rivers.

The Batesville in the Wyandotte quadrangle is, in part, overlain with apparent conformity by the Fayetteville Shale, and in part, overlain unconformably by the Krebs Group of Pennsylvanian age. In much of the subsurface in the mining field, and also in surface outcrops at, and a couple of miles south of Miami, the Batesville was removed by erosion at the unconformity, so that the Krebs Group rests unconformably on stratigraphic units underlying the Batesville (pl. 2).

Fayetteville Shale

The Fayetteville Shale crops out in a broad northeast-southwest belt of poor exposures on the prairie between Narcissa and the edge of the quadrangle west of Afton and in outlying patches to the east on the flat drainage divide north and southeast of Fairland. A second northeast-trending belt of extensive, though poor, outcrop is along the southeast flank of the Horse Creek anticline from Crossland to Duck Creek southwest of Cleora. The only mapped occurrences east of the Spring-Neosho River system are in small sinkholes, several on the northwest side of Buffalo Creek opposite the mouth of Beeman Branch and one on the north edge of the quadrangle north of Spring City. The unit is, however, probably included in the jumbled Chester strata in places along the Seneca graben block.

A segment of the Fayetteville is exposed underground in the Blue Goose No. 1 mine southwest of Cardin, in a drift that crosses the deepest part of the Miami trough. The trough here is a graben block about 300 feet wide (pl. 9). A subsidiary fault of small normal displacement, striking parallel to the graben, is exposed in the drift, halfway between the two bounding faults. Fayetteville strata are confined to the southeast side of the subsidiary fault, with the limestone and sandstone of the Batesville on the northwest side.

The Fayetteville may be present locally in a few other structural basins in the mining field, but it has not been detected. Truncation of Chester strata by the unconformity at the base of the Pennsylvanian shale is known to be widespread in the mining field, and any additional remnant of Fayetteville in the subsurface would be of small areal extent at best.

Thickness

Near Crossland the Fayetteville Shale is about 70 feet thick, presumably as measured in a well.[5] In a poorly exposed section just outside the Wyandotte quadrangle boundary southwest of Afton, the thickness is about 55 feet. Elsewhere, only parts of the unit are exposed; the full thickness, therefore, cannot be measured. If all the fossiliferous limestone associated with shale in the underground exposures at the Blue Goose No. 1 mine is Fayetteville, the thickness cut here is 30 feet; probably neither the base nor top is exposed.

Character

The Fayetteville in its outcrop areas consists of black, dark-gray, bluish-gray, and greenish-gray fissile shale containing limestone in at least one zone that is as much as 9 feet thick, and containing limestone and limy shale in thinner bands, in part nodular, interbedded in parts of the shale. The limestones and some of the limy shale bands are notably fossiliferous. Exposures are not good enough to determine whether the thick limestone is at the same horizon in different places, but the presence of only one such zone in each of several different areas suggests that the same stratigraphic unit is involved, lying perhaps near the middle of the Fayetteville. The shale contains ironstone lenses and scattered limonitic concretionary shells that enclosed egg-shaped masses of shale as large as 2 inches across. The shale weathers buff or brown. Locally, it grades to fine-grained or shaly sandstone near or at the top of the formation.

The limestone is commonly crinoidal, or oolitic with the spherules elongated. However, the thinner beds may be, in part, fine grained and semicrystalline. The coarser grained limestones and oolites in the belt of outcrops just beyond the quadrangle boundary southwest of Afton are brown and blue gray, some with a purplish tinge; but in the belt southeast of the Horse Creek anticline, they are commonly dark, have a black matrix, and may be bituminous. The fine-grained limestone is blue gray to dark gray. The limestones may contain thin films of limonite on joint cracks, and commonly weather to yellow or buff colors, but some of the fine-grained ledges may locally weather to nearly white surfaces.

The underground exposures in the Blue Goose No. 1 mine lie along 150 feet of drift that crosses a flat, nearly symmetrical syncline between 2 faults within the Miami trough system. Maximum dips of 25°–30° occur on each of the two limbs near the delimiting faults. Possibly the basal 6 feet of strata in the fault block would be classed as Batesville, if an unbroken stratigraphic sequence were accessible for determination of the stratigraphic succession. The limestones here are gray, dark gray, and brownish gray, and some of them are extremely fossiliferous. The section is as follows:

Top of section, at roof of drift on axis of syncline.

	Feet
1. Shale, dark-gray with indefinite greenish seams and zones. Exposed below roof	2
2. Oolite, ooliths rounded, oval, and oblong, in uneven sizes; 3-in. basal intraformational conglomerate has rounded, oval, cigar-shaped, and half-cigar-shaped pebbles in cross section, lying parallel to bedding; the largest pebble is a half cigar 2½ in. long and ¾ in. in diameter; pebbles also composed of oolite, but darker colored, and ooliths have different size and packing density compared with those in surrounding matrix	1
3. Limestone, gray to brownish-gray, rather fine-grained, fossiliferous; numerous greenish shale partings	1
4. Oolite, gray, massive, slightly fossiliferous; ooliths rounded, oval, oblong	1

[5] From a cryptic note in C. E. Siebenthal's handwriting on his manuscript map of pl. 2.

	Feet
5. Shale, greenish, unfossiliferous	1¼
6. Limestone and shale in about equal amounts; limestone gray, somewhat pyritic, fine grained except for abundant fossils, including crinoid stems, which make it appear medium to coarse grained; poorly and indifferently oolitic, but some purer limestone is good oolite with rounded and oval ooliths; poorly, discontinuously, and lenticularly bedded with irregular and indefinite shale partings. Shale greenish to gray, in part limy, highly fossiliferous; comes and goes lenticularly in the limestone in laterally variable proportions. Fossils include *Archimedes;* other bryozoa common	8
7. Shale, limy, greenish, highly fossiliferous; bryozoa especially abundant; local thin lenticular nodules of limestone	1
8. Shale, very dark gray ("black"), unfossiliferous	4
9. Limestone, dark-gray, pyritic; probably has fine-grained matrix but abundant fossils make it appear medium to coarse grained. Crinoid stem fragments, dark-flesh colored to very dark gray, not abundant but may be large and conspicuous	¾–1
10. Shale, very dark gray ("black")	1½
11. Limestone, gray, coarse grained with some medium grained at top, apparently very fossiliferous; contains lenticular to oval "pebbles" which may be organic, 1–2 in. long, of dark-gray fine-grained limestone lying parallel to bedding	2
12. Limestone, dark-gray, medium- to rather fine-grained; conspicuously fossiliferous in upper few inches; rather massively bedded	4½
13. Alternating limestone (predominant) and green shale; limestone gray, fine-grained, but has rather well-defined bands and scattered single crystals of coarser crinoidal material	1½
Base of section at normal fault probably of small throw, striking northeast and dipping 60° SE.	
Total	29½

The lithology, stratigraphic sequence, and abundance of fossils are very similar to parts of the Fayetteville revealed in outcrops farther south, though this underground section has relatively more limestone. Fossils collected from beds 6, 7, and 9 of the section yielded about 20 forms identifiable as to genus or species. Concerning these, Mackenzie Gordon, Jr. (written commun., 1965), comments as follows:

> The presence of the recognizable species such as *Orthotetes kaskaskiensis* (McChesney), *Ovatia minor* (Snider), *Punctospirifer transversus* (McChesney) ?, *Reticulariina spinosa* (Meek and Worthen), *Torynifer setiger* (Hall), and *Composita subquadrata* (Hall), all in association, add up to a definite Chester age for this fauna. The presence of numerous well-preserved bryozoans and an undescribed species of *Schuchertella* that is common in the Fayetteville Shale in this region identifies the beds as belonging in the Fayetteville Shale.

The immediately underlying Batesville, Hindsville, and possibly Quapaw strata are not exposed in the structural block containing the Fayetteville; but in the other half of the Miami graben block, lying to the northwest just across the small fault indicated at the base of the section given, roughly 88 feet of these strata occurs between the top of the Boone and the highest stratigraphic horizon cut in this part of the drift. Only the upper 22 feet of this section, predominantly limestone, is exposed in the drift; the remainder was cut in a core hole drilled from the floor of the drift. There may be an additional unknown thickness of Batesville above the roof of the drift in this block.

Correlation and Age

The limestones and shaly limestones contain abundant fossils of Chester age which correlate with those of the type Fayetteville. The type locality is in the Fayetteville quadrangle, Arkansas, which lies southeast of the Wyandotte quadrangle and has a common corner, though the closest mapped outcrops of the Fayetteville Shale in the two quadrangles are 35 miles apart. The formation has the same stratigraphic position in the two areas; but the lithology has changed to a more limy facies in the Wyandotte quadrangle, and the Wedington Sandstone Member, present in the type area, is missing in the Wyandotte quadrangle, or, at best, is represented only by an ill-defined sandy zone in the upper part of the shale. Nevertheless, there is enough lithologic similarity between the two areas to warrant correlation even independently of the fossil evidence.

Gordon (written commun., 1965) has supplied the following comments on the faunal content:

> The fauna of the Fayetteville Shale in the Wyandotte quadrangle consists of about 180 species of invertebrates, as determined from 32 fossil collections. The bryozoans are the most abundant, both in numbers of individuals and species, but are fairly closely followed by the branchiopods and mollusks. Also relatively abundant are the ostracodes, of which more than 30 species have been recognized. Less common are the echinoderms, corals, endothyroid protozoans, and trilobites.
>
> Most of the Fayetteville brachiopods are typical Chester species, the majority of which are found in the Hindsville Limestone and the Batesville Sandstone. Particularly characteristic of the Fayetteville and not found in the earlier Chester beds are *Schuchertella* n. sp. and *Stenocisma explanata* (McChesney). *Inflatia adairensis* (Drake), relatively rare in the Hindsville and Batesville, is common in the Fayetteville.

Stratigraphic Relations

The Fayetteville Shale overlies the Batesville Sandstone, or equivalent strata in the Hindsville Limestone where the sandstone is absent, with apparent conformity. It is overlain unconformably by the Hale Formation of Early Pennsylvanian age in the belt of outcrop along the southeast side of the Horse Creek anticline. In the outcrop from Afton to Narcissa it is overlain by

the Krebs Group of somewhat later Pennsylvanian age. An unconformity of the base of the Krebs cuts out the Fayetteville north of the Narcissa area, except for a remnant preserved underground in the Picher field in a graben block of pre-Krebs age.

PENNSYLVANIAN SYSTEM

Although several formations of the Pennsylvanian System crop out in the Wyandotte quadrangle, only the lower part of the thick section of rocks belonging to this system in the midcontinent area is present. Two series are represented, of which the older crops out in only a small area and owes its preservation to very special conditions in the geologic history of the region. The younger series covers an extensive area of the quadrangle.

MORROW SERIES

HALE SERIES

The Hale Formation was formerly exposed in a hill within the Seneca graben block on the west side of the Neosho River, 2 miles southwest of Sailboat Bridge. Except for a few feet at the top of the hill which now forms a small island at low water stages, this locality is flooded by Lake O' the Cherokees, as are other areas of Hale to the northeast and southwest along the graben block. Sandstone float from the basal part of the formation caps the cemetery knoll southwest of Bernice, and is more sparingly scattered over fields and pastures at several other localities along the southeast side of the Horse Creek anticline from Crossland southwest to Duck Creek. A small outcrop of the sandstone is exposed in a roadside grading ditch near the section corner at the southwest corner of sec. 17, T. 25 N., R. 23 E.

THICKNESS

The section as measured by Siebenthal on the west side of the Neosho southwest of Sailboat Bridge aggregates 83 feet from the base of the lowest sandstone to the top of the hill, with no overlying formation to delimit the top. The basal 22 feet of this section could be either Fayetteville or Hale, for fossils have not been found in it; but the next higher sandstone has fossil plants which, although indeterminate in themselves, surely indicate affinity with the overlying strata. It is believed that the basal 22 feet, in which the sandstone is very similar to that in overlying beds, should also be classed as Hale, and it is so considered in the following discussion.

CHARACTER

The Hale Formation in its lower 41 feet at the Neosho River locality consists of sandstone with interspersed covered intervals that are presumably shale. The sandstone is chiefly massive, but in part flaggy, rather fine grained, soft, and brown to reddish brown, with a 1-foot bed of ferruginous sandstone or sandy iron ore. The sandstone contains abundant plant fragments as much as 3 feet long, either coaly on freshly broken exposures or as impressions on the weathered surface. Some of the sandstone is colored black from inclusion of finely carbonaceous or bituminous material. What little shale that cropped out with the sandstone is gray to black, and weathers to paper-thin flakes.

The upper 42 feet consists of interbedded limestone and calcareous sandstone, alternating in beds commonly 5–10 feet thick, with local lateral gradation from one to the other. The limestone beds are coarse-grained organic, in part coquina, and include some oolite. The sandstone is rather fine-grained and leaches to reddish brown. Both are heavily pigmented with dark bituminous material, possibly oily.

As this section is unique, Siebenthal's detailed description is given, which somewhat overemphasizes the blackness or "oiliness" of the section.

Section of Hale Formation formerly exposed on west side of Neosho River, 2 miles southwest of Sailboat Bridge

[Measured by C. E. Siebenthal]

Top of hill.

	Feet
Black fossiliferous crinoidal limestone	1
Concealed	4
Fossiliferous calcareous sandstone	1¼
Oily black crinoidal limestone, basal part concealed	5
Oily black crinoidal limestone, with oolite	10
Black calcareous sandstone	5
Heavy black limestone, granular, with oolite	5
Black limy sandstone, becomes more calcareous to west, makes good limestone; fossiliferous	11
Concealed	5
Coarse-grained black sandstone	1
Concealed	3
Ferruginous sandstone or sandy iron ore	1
Flaggy black sandstone, similar to mass below; plant remains	9
Concealed	10
Oily looking but dry, rather fine-grained heavy sandstone	12
Total	83

Concealed shale and thin blue limestone of the Fayetteville.

The sandstone found in remnants along the southeast side of the Horse Creek anticline between Crossland and Duck Creek is less "oily" than the sandstone in the lower part of this section, but it is otherwise similar, and contains the coaly material and plant impressions.

CORRELATION AND AGE

Fossils collected by Siebenthal and Roundy from limestones and limy sandstones in the upper half of the

section described above were considered by G. H. Girty to be of Morrow age. Gordon (written commun., 1965) has made the following comments:

> Two collections, from near the middle and from the top few feet of the section measured by Siebenthal, have yielded 23 species of invertebrates including remains of brachiopods, mollusks, corals, bryozoans, echinoderms, and trilobites in descending order of number of species. Species identified by G. H. Girty and indicative of Pennsylvanian (Morrow) age are *Schizophoria altirostris* (Mather), *"Productus" welleri* Mather, *Spirifer* aff. *S. rockymontanus* Marcou, *Condrathyris perplexa* (Swallow), *Hustedia* aff. *H. miseri* Mather, *Cleiothyridina orbicularis* (McChesney), and *Pseudomonotis precursor* Mather. This is the northernmost outcrop occurrence of fossils of Morrow age in Oklahoma.

Plant fragments collected from sandstones in the basal half of Siebenthal's section are not well enough preserved to be identifiable for age determination. Other plant remains collected by us from the basal sandstone in the railroad cut a quarter of a mile southwest of Bernice contain impressions of a calamarian stem, probably *Calamites*, and impressions of probable cordaitean leaves; the latter indicate a Pennsylvanian age, though the specimens are too poor for more precise determination (Sergius Mamay, written commun., 1964).

Rocks of the Morrow Series in Oklahoma are divided into the Hale Formation below and the Bloyd Formation above. The Bloyd is present in the general belt of Pennsylvanian outcrop east of Muskogee to the Arkansas State line, but northward is truncated at about the Mayes-Wagoner County line by the unconformity at the base of the overlying Atoka Formation (Huffman, 1958, p. 80; C. A. Moore, 1947, p. 46, pl. 13). The underlying Hale Formation is also truncated farther northward by unconformity, but it has been traced by Huffman and others to the general vicinity of Strang. One of the largest remnants is preserved in the Seneca graben block just southwest of the Neosho River at Strang (Huffman, 1958, pl. 1). In this remnant the Hale contains sandstone in the lower 31 feet, overlain by 8 feet of sandy limestone (Huffman, 1958, section 158, p. 265). The section preserved in the Seneca graben, 25 miles farther to the northeast, as described in this report, is similar except that the basal sandstone contains some interbedded shale, and there is considerably more limestone present above the sandstone. This section is therefore correlated with the Hale. It shows stratigraphic similarities to the Hale Formation at localities 50 miles to the southeast in Washington County, Ark. (Henbest, 1953). These similarities include the presence of argillaceous material in the lower part and the presence of a calcareous upper part that includes fossiliferous crinoidal limestones and some oolites in addition to calcareous sandstones.

At the time that Siebenthal mapped the sandstone occurrences southeast of the Horse Creek anticline, he presumed that they correlated with the Wedington Sandstone Member of the Fayetteville Shale (Siebenthal, 1908, p. 190; 1915, p. 28). However, the sandstone bears little resemblance to the Wedington of the type locality, 40 miles southeast in Arkansas, nor to the isolated small outlier of mapped Wedington lying 26 miles southeast of the outcrops under discussion (see Slocum, 1955, p. 20, 38). There is no Wedington along the main belt of Fayetteville outcrops to the southwest for many miles (Snider, 1915, p. 35–37). The fossil plants indicate that the sandstone along the Horse Creek anticline is of Pennsylvanian age. Close proximity and similarity of this sandstone to that in the lower part of the section formerly exposed in the Seneca graben suggest correlation with the Hale Formation rather than with later Pennsylvanian sandstones.

Stratigraphic Relations

The Hale Formation rests unconformably on underlying formations in areas to the southwest of the Wyandotte quadrangle (Huffman, 1958, p. 76). Its contact with the underlying Fayetteville Shale is nowhere exposed in the quadrangle, but the relative thinness of the Fayetteville is consistent with the known regional truncation of underlying strata by the Hale, the hiatus increasing gradually to the northeast.

No overlying strata are in contact with the Hale in the Wyandotte quadrangle. The next younger formation, the Hartshorne at the base of the Krebs Group, crops out several miles northwest of the Hale exposures, but as it is unconformable on the underlying strata, it has there overlapped the Hale to rest on the Fayetteville. Local preservation of the Hale beneath the unconformity is possibly due to pre-Krebs downwarping and graben faulting along the asymmetric Horse Creek anticline and Seneca graben.

Scattered remnants of probable Atoka Formation, whose stratigraphic position is between the Hale and Hartshorne Formations, have been mapped in the area just west of the southern part of the Wyandotte quadrangle (Huffman, 1958, p. 84, pl. 1; Miser, 1954). These are outliers beyond the northern edge of the Atoka, which is truncated and overlapped several miles southwest of the Wyandotte quadrangle by the unconformity at the base of the Krebs Group. According to unpublished work by Robert B. Branson,[6] these outliers are channel sandstones cut into the underlying Fayetteville Shale. The Atoka is unconformable on underlying for-

[6] Robert B. Branson, 1952, Geology of the Vinita-Ketchum area, Craig County, Okla.: Oklahoma Univ., unpub. M.S. thesis.

mations (Huffman, 1958, p. 85). If remnants of Atoka similar to those just to the west ever overlay the Hale in the Wyandotte quadrangle, all evidence has been removed by erosion that could have been either pre-Hartshorne or relatively recent.

DES MOINES SERIES

KREBS GROUP

All rocks of Des Moines age in the Wyandotte quadrangle were formerly classed as Cherokee Shale, which includes 2 or 3 sandstone members. In recent years, several formations that have long been recognized in the McAlester basin in Oklahoma have been mapped northeastward along the border of the Ozark uplift to the Kansas line. Five of these formations are included within the limits of the Cherokee, but only the basal four crop out in the Wyandotte quadrangle. C. C. Branson (1955), following Oakes (1953), has used the term Krebs Group for these four formations, comprising from the base upward, the Hartshorne, McAlester, Savanna, and Boggy Formations. In the Wyandotte quadrangle they are predominantly shales with a few sandstone, siltstone, limestone, and coal beds marking definite horizons. They have been mapped in Ottawa County by Branson (in Reed and others 1955, pl. 1). Table 3 gives the formations and members of the Krebs Group, with approximate thicknesses.

TABLE 3.—*Formations and members of the Krebs group in the Picher mining field*

[Modified from C. C. Branson, 1955]

	Approximate thickness[1] *(feet)*
Top.	
Boggy Formation (upper part eroded): Bluejacket Sandstone Member (at base)	60
Savanna Shale	120
Unnamed sandstone member (51 ft below top)	12
Doneley Limestone Member of Branson (1954) (15 ft below the sandstone)	4
Rowe coal (2½ ft below Doneley limestone Member)	⅓
McAlester Shale	30–50
Unnamed shale member	30
Warner Sandstone Member (at base)	0–20
Hartshorne Formation	0–50
Riverton coal (at top)	0–⅔
Unnamed coal (just above base)	¼

[1] Thicknesses only in part from Branson.

The mapping of the undivided Krebs Group, as shown on plate 2 of this report, is virtually that of Siebenthal and Mesler. A sandy zone which they had mapped for a few miles south of Narcissa, and which they had termed "Wedington," is here placed in the Krebs, following C. C. Branson (1955). The mapping is more generalized than that of Branson, and evidently some faults cutting the basal Krebs contact northeast of Narcissa, as well as others entirely within Krebs outcrops northwest and southwest of Narcissa, were missed.

The shales of the Krebs Group underlie the northwest corner of the quadrangle. The base of the group follows an irregular northeast line, extending from a mile west of Afton through Quapaw, and forms a pronounced reentrant to the northwest along the Neosho River at Miami. The shale is the surface rock over the greater part of the mining field. It also occurs, generally with associated sandstone facies, in the Seneca graben block near Seneca and Racine and in scattered sinkholes on the upland east of the Spring and Neosho Rivers north of the Elk River.

THICKNESS

The maximum thickness of the Krebs Group in the quadrangle is found along the Miami trough through the Picher mining field, and amounts to about 275 feet. Higher strata of the group crop out west of the quadrangle, beyond the mineralized area. The highest stratigraphic unit cropping out within the quadrangle is the Bluejacket Sandstone Member at the base of the Boggy Formation. It is commonly 15–50 feet thick, but is 60 feet thick in one drill hole on the Scammon Hill tract north of Commerce. Its base is 190 feet above the base of the Krebs Group in this hole. Weidman (1932, p. 25) quotes thicknesses of 197–248 feet for the interval below the Bluejacket in other holes on this tract, but it is possible that the interval may be structurally thickened somewhat in some of these holes by crumpling and folding along the trough axis. On the Blue Mound in Kansas, just north of the State line at Picher, the interval in 3 drill holes is 195–210 feet (Weidman, 1932, p. 24). Perhaps a good average for the stratigraphic interval in the mining field would be 190–210 feet.

The Warner Sandstone Member at the base of the McAlester Shale, formerly called the Little Cabin Sandstone Member of the Cherokee (Weidman, 1932, p. 23), is a fairly good marker bed in the lower part of the sequence. It ranges in thickness from 0 to 20 feet, but reaches local maxima of 30 feet in drill holes on the Federal Jarrett and Robinson mining tracts where it may have been thickened by structural adjustments on sharp synclinal folds. In undeformed ground along the Lucky Jew-Black Eagle line, the Warner is uniformly 20 feet thick in numerous drill holes. Elsewhere, it is rarely more than 15 feet thick, and averages peraps 10 feet, with many occurrences in the 5- to 7-foot range. Its

base is commonly 30–50 feet above the base of the Krebs Group, but on much of the Federal Jarrett, the Robinson, and along the Lucky Jew-Black Eagle line, the interval ranges from 12 to 30 feet, and in a single drill hole in which the sandstone is recorded on the Early Bird tract (SW¼NW¼ sec. 12), this interval is 15 feet. In the mineralized area just north of the State line southwest of Melrose, Kans., what is believed to be the Warner Sandstone Member was cut in numerous drill holes, but not all. It is as much as 15 feet thick. It lies at the base of the Krebs in most places, but is as much as 20 feet above the base in a few drill holes. Pierce and Courtier (1938) record the interval below the Warner as ranging from 0 to 30 feet but normally 15 to 20 feet in Cherokee County, Kans., just north of the Wyandotte quadrangle.

Character

The Krebs Group consists of black and dark-gray fissile shales with a few gray seams, and three sandstone members. These include the Warner, Bluejacket, and an unnamed third sandstone member which C. C. Branson describes in the Savanna Shale (table 3). Other sandstone or siltstone beds occur sporadically, with no demonstrated continuity, the base of the group being particularly prone to sandiness. A dark carbonaceous fossiliferous limestone about 4 feet thick in the Savanna Shale, 15 feet below the unnamed sandstone, has been named the Doneley Limestone Member by Branson (1954), and there are scattered thin calcareous layers at other levels, as well as thin layers of clay ironstone concretions. Thin seams of coal occurring at several levels are mostly less than a foot thick. However, on the Blue Mound, a prominent hill just north of the State line at Picher, a bed of coal lying 4 feet below the base of the Bluejacket Sandstone Member is 14–16 inches thick; 10 feet lower in the section is another coal seam, 12 inches thick.

C. C. Branson (1955, p. 64–67) has given detailed sections of the Hartshorne and McAlester Formations.

Basal conglomerate.—The basal shale of the Krebs contains pebbles of chert at numerous places. At the cavein over the Discard mine, these pebbles are 1–3 inches across, irregularly rounded, and are confined to a very thin basal layer. They contain poorly preserved casts of crinoid stems and poorly defined blebs of massive pyrite. Weidman (1932, p. 22–23) describes a similar occurrence in a shaft on the Scammon Hill tract and quotes other occurrences in the mining field. The pebbles may be derived from the Boone, but enough chert is in the Hindsville to make this an even more probable source, particularly where the Hindsville is the subjacent rock, as at the Discard mine. C. C. Branson (1955, p. 64) describes a surface section in which the cobbles in the basal conglomerate are, in part, limestone.

Not all sections show a conglomerate at the base of the Krebs. An underground exposure in the Roanoke mine shows black shale at the base, but a 1-foot seam of gray fine-grained shaly sandstone is bedded in the shale just 1 foot above the base. The section in the Central mine shows at the base a dark-gray siltstone, 1 foot thick, in which the silt tends to form paper-thin laminae in the dark shale.

Warner Sandstone Member of McAlester Shale.—The outcrop of the Warner Sandstone Member in Ottawa County has been mapped by Weidman (1932, pl. 1) and C. C. Branson (in Reed and others, 1955, pl. 1). Additional information on the sandstone has been obtained from prospect drilling; but owing to the drilling habits in the mining field whereby all cuttings above the limestone or chert are discarded and usually not logged or only incompletely logged, much information on the Warner Sandstone Member is lost. The fact that it does not appear in the driller's log does not necessarily mean that it is not present. Particularly where it is shaly, it is likely to be ignored. However, enough information was gleaned from the drilling examined by us to indicate that the sandstone is quite variable in thickness and is locally discontinuous, at least to the extent that the driller has reported "no sandstone" in certain drill holes whereas he has logged it, or the cuttings have been saved, from other holes.

The Warner Sandstone Member varies from fine to predominantly coarse grained and is pale gray, pale brown, or pale green where fresh, but weathers buff, brown, or reddish brown. The sand grains are commonly elongate and angular, or may show partial rounding, but they grade to well rounded. Much of the sandstone is micaceous. It is commonly crossbedded. In places the unit contains gray shale, apparently interbedded, or it may grade to sandy shale.

Bluejacket Sandstone Member of Boggy Formation.—As mapped by Weidman (1932, pl. 1), the Bluejacket Sandstone Member forms a series of disconnected outcrops along the axis of the Miami trough, from Commerce northeast to the Kansas Explorations Jarrett tract 2 miles north of Picher; it also caps the upper 25 feet of the Blue Mound in Kansas a short distance east of the Miami trough. Other outcrops cap two hills near the Craig County line, 4–6 miles west of Miami. Pierce and Courtier (1938, pl. 1) have mapped the Bluejacket widely in Cherokee County, Kans., adjoining the area in which it was mapped by Weidman. This sandstone is light gray where fresh, but weathers yellow, buff, and reddish brown.

Southwest of Melrose, Kans., about half a mile from the State line, prospect drilling in the SW¼ sec. 11, T. 35 S., R. 22 E., cut the Bluejacket Sandstone Member in several holes. According to the driller's logs, it is commonly 40–52 feet thick, but may be as thin as 15 feet. Its base lies 150–170 feet above the base of the Krebs Group here. Although the Warner Sandstone Member is recorded in only one of the drill holes in which the Bluejacket is shown, this and other drill holes in the general vicinity indicate that, where present, the Warner lies at, or close to, the base of the Krebs. Thus, although the Bluejacket is about 40 feet closer to the base of the Krebs than at Picher, the reduction of the interval takes place largely below the Warner Sandstone Member, so that the interval between the two sandstones is about the same as at Picher and in other parts of Cherokee County, Kans. (Pierce and Courtier, 1938, p. 25). One of the drill holes southwest of Melrose cut coal just below the Bluejacket, which may correlate with one of the coal seams previously mentioned (p. 69) below the Bluejacket Sandstone Member on the Blue Mound, or with the Columbus coal in other parts of Cherokee County, Kans. (Pierce and Courtier, 1938, p. 64), or with the Drywood coal in Craig County, Okla. (Reed and others 1955, p. 68).

Correlation and Age of Krebs Group

The Warner Sandstone Member of the McAlester Formation, the Bluejacket Sandstone Member of the Boggy Formation, and intervening strata have been mapped in Haskell County, Okla., by Oakes and Knechtel (1948), across parts of McIntosh and Muskogee Counties by Wilson and Newell (1937), and thence into Ottawa County by C. C. Branson (in Reed and others, 1955, p. 66–68). These Krebs subunits in the Picher mining field are thus tied by direct mapping to the standard Des Moines section of Oklahoma in the McAlester basin. Although not similarly mapped through, the two thin coal seams at the top and bottom of the Hartshorne (table 3) are tentatively correlated by C. C. Branson (in Reed and others 1955, p. 64, 65) with the Upper and Lower Hartshorne coals of the standard section. From a study of the fossil flora in the McAlester basin, Read (Hendricks and Read, 1934, p. 1055, footnote) correlates the formations from Hartshorne to Boggy inclusive with the basal part of the Allegheny Series (Clarion) of the western Pennsylvania coal fields.

In the erosion interval preceding deposition of the Krebs, caves were formed by solution of the limestone in certain favorable zones of the Boone in the Picher area. These caves were later filled by the Pennsylvanian black shale, in places with a little sandstone, presumably during the initial invasion of the land by sediment-bearing waters at the beginning of Pennsylvanian deposition. Many such filled caves are revealed in mine workings in the Joplin Limestone Member (M bed). They are commonly a few tens or hundreds of feet across and only a few feet high, though locally as much as 10 feet high. The shales are well laminated, generally undeformed, and contain carbonized plant remains, including wood fragments and seeds. The seeds are sharp angled and unworn, and were washed into the caves along with the enclosing mud while still woody, probably a short time after their growth. A collection from the Kenoyer mine (NE¼SW¼ sec. 20) was examined by Charles B. Read, of the Geological Survey (Oct. 8, 1934), who reports as follows:

> The collection of fossil plants from a sink in the Evans-Wallower No. 8 [=Kenoyer] Mine, Tri-State district, includes the species listed below:
>
> *Cordaites* cf. *C. communis* Lesquereux (fragments)
> *Cordaicarpus lineatus* Lesquereux (abundant seeds)
> *Rhabdocarpus* sp. indet. (1 fragment)
> *Calamites* sp. indet. (1 fragment)
>
> This aggregation of plants furnishes very little information of correlative value except indirectly. The collection is, of course, Pennsylvanian in age. Furthermore, the abundant seeds (*Cordaicarpus lineatus* Lesquereux) are middle Allegheny in their Appalachian occurrence, and in the midcontinent region are known only from a sinkhole in the Boone near Joplin, Mo. There these same seeds, partly replaced by marcasite as in the present case, are associated with a flora of 32 species which is clearly upper or middle Cherokee. The meager information at hand thus indicates a similar age for the shales in the Evans-Wallower mine, if there has been no reworking of these plants. The rather abundant fragments of mineral charcoal do suggest that there has been some reworking of materials originally deposited in a coal bed. There are several coals in the Cherokee which might furnish this charcoal, and such redeposition might take place in nearly contemporaneous beds. The seeds show no evidence of transportation, however, the pointed apices of most of them being well preserved.

Stratigraphic Relations

The Krebs Group rests unconformably on underlying formations. From west of Afton to Narcissa it lies on the Fayetteville Shale. On the outcrop, northeast of Narcissa to the Kansas line, it lies chiefly on the Batesville Sanstone, but locally, its base cuts below the Batesville to rest on the Hindsville Limestone. In the subsurface of the main mining field it rests in places on the Batesville, chiefly where the latter has been preserved in certain structural basins. At one place in the Miami trough it rests on a downfaulted remnant of the Fayetteville Shale. At most places in the mining field, however, it lies on the Hindsville Limestone or perhaps on the Quapaw Limestone. In a few structurally high

places west of the Miami trough and Bendelari monocline, all Chester and Quapaw strata were eroded, so that the Krebs rests on the Boone. In the mineralized area southwest of Melrose, Kans., the Krebs rests chiefly on a truncated remnant of the Hindsville or possibly Quapaw; but locally, it also rests on the Boone. In this area the Warner Sandstone Member, which normally lies above the base of the Krebs Group, has apparently cut downward by overlap of the basal Hartshorne unit; and although incompleteness and imperfections of the drilling records leave some gaps in the supporting evidence, it is believed that the sandstone at the base of the Krebs in much of the Melrose area is the Warner Sandstone Member. If so, the hiatus represented in the unconformity at the base of the Pennsylvanian strata is greatest in this area.

In addition to the normal erosion of earlier strata to a peneplain surface in the interval preceding Pennsylvanian sedimentation, subsurfaces leaching and cave formation in limestone strata of the bedrock were intensified. The absence of any known erosion below the peneplain on which the initial sediments of the Pennsylvanian were deposited suggests that the shale-filled caves were formed below the zone of ground-water saturation.

Although most of the underground caves were filled by the black Krebs shales, locally, some of them had been filled by a residual shale prior to the invasion of the Krebs. At one place on the upper level of the Central mine, such a cave in the lower part of the Batesville strata is filled with a gray waxy mudstone containing some sand grains and small fragments, lenses, and rounded balls of light- to dark-green clay, all probably residual from leaching of limestone of the Batesville. At the top of the cave, where it is 7 feet wide at the plane of the section, are blocks of limestone from a horizon that must be stratigraphically higher than anything preserved elsewhere below the basal Krebs unconformity, to judge from the exposure on the left side of the cave (fig. 13). Thus, the cave must have formed, evidently along a small fault break, and filled at a time when there were limestone walls above the present cave remnant that could furnish the limestone blocks and residual clay to the cave filling. Pieces of fossil wood found in the gray mudstone filling 12–15 feet below the unconformity were introduced at the same time, and are older than basal Krebs, though the fossil wood at another place appears to be part of a tree-root system that could have been nearly contemporaneous with basal Krebs. At some time after planation of the unconformity surface and deposition of the basal Krebs, the thin siltstone at its base slumped a few inches into the top of the cave, but the black shale immediately overlying it did not break, showing that the top of the cave was stabilized just after the beginning of the Pennsylvanian deposition. At other places in the neighborhood of this cave, however, the black shale broke and slumped into underlying caves.

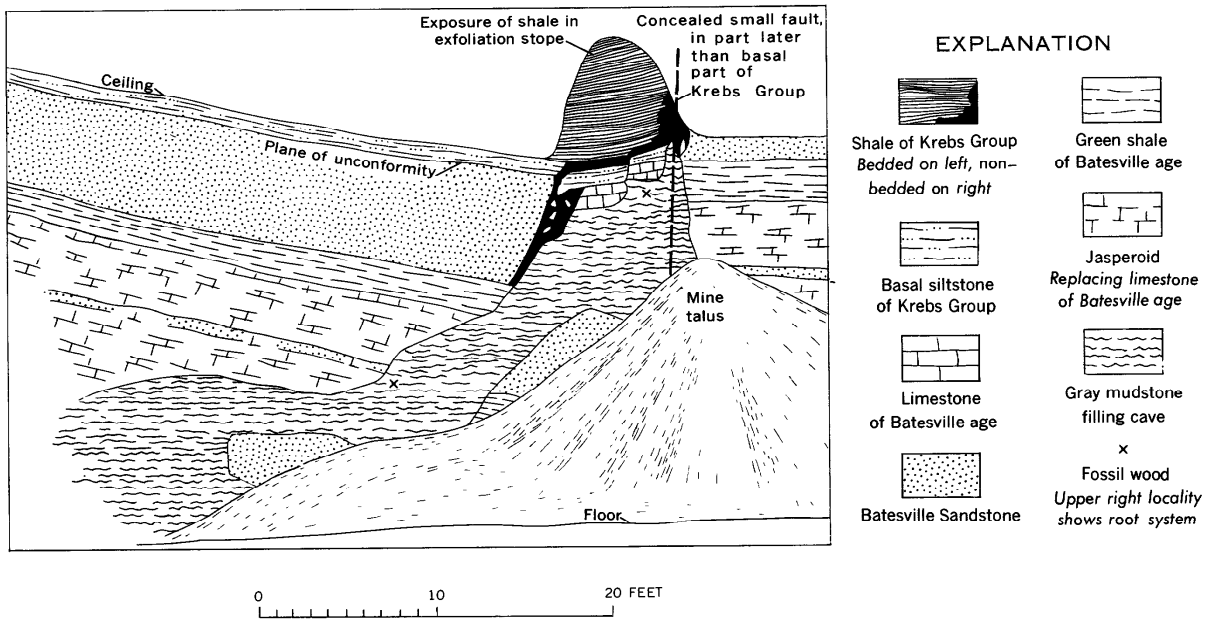

FIGURE 13.—Sketch of filled cave in Batesville Sandstone on upper level of Central mine.

STRUCTURE

General Features

The Wyandotte quadrangle is on the northwest flank of the Ozark uplift, though near the southwest end of this broad flat elliptic dome. The rocks of the quadrangle are for the most part nearly flat, but on a regional average they dip northwestward about 25 feet per mile in the northwest half of the quadrangle, and about 20 feet per mile in the southeast half. Locally, they may dip in other directions, which disguises the regional dip. In a few places, sharply defined structural features may be accompanied by appreciable dips. One of these features, the Horse Creek anticline, which is so asymmetric as to be, in effect, a monocline, reverses part of the displacement caused by the regional dip. Thus, averaged roughly across the whole quadrangle, the northwesterly regional dip amounts to about 19 feet per mile. Two pronounced crustal breaks, the Seneca graben and Miami trough, trend northeast across the qudrangle, and are expressed in the surface rocks as graben blocks or narrow synclinal sags. Features of intermediate size revealed by detailed structural mapping in the Picher mining field include the Bendelari monocline that trends northwest across the Miami trough, and several pronounced synclinal basins as much as 1 mile in longest dimension but more commonly confined within a 40-acre (a quarter of a mile square) tract.

In addition to the features that have been produced by tectonic stresses, many local features of smaller size have been produced by slumpage following the dissolving action of ground water and mineralizing water on carbonate rocks. Although the process causing the local features is quite distinct from that causing the larger ones, the two are commonly closely associated in space; hence, many structural features have a hybrid origin, and it may be difficult to appraise the relative effects of the two causitive processes. Some of the smaller synclinal basins that are nearly round in plan are particularly puzzling.

Surprisingly, few tectonic faults have been recognized outside the Seneca and Miami zones of crustal break. Two faults, however, are associated with the Horse Creek anticline, though apparently of small throw. Probably more detailed mapping of the stratigraphic units as recognized in the present report would reveal a few more faults, comparable to those revealed by more detailed mapping of the Pennsylvanian (Reed and others, 1955, pl. 1). It is not probable, however, that any large faults have been overlooked. Those found by detailed underground mapping in the Picher field, except where related to the Miami trough block, are short and of small displacement, and probably most die out in the section a short distance above or below the mine workings. Many such faults are apparently due as much to solution slumpage as to tectonic stresses.

Horse Creek Anticline

The Horse Creek anticline curves in a general northeasterly direction across the Wyandotte quadrangle a short distance south of its center, from Cleora on the southwest, through the high bluffs on the Neosho River a couple of miles north of the Delaware County line, and continuing more easterly to cross Buffalo Creek about 3 miles above Tiff City. The trend is about N. 55° E. at the southwest end near Cleora, and about N. 85° E. from the Neosho River to Buffalo Creek. The anticline was named by Siebenthal (1908, p. 198) for Horse Creek, which flows across the structure southeast of Afton.

The fold is asymmetric, with gentle dip, 1° or less, on the northwest flank and, as described by Siebenthal (1908, p. 198), a steeper dip, 5°–18°, on the southeast flank and is, in effect, a monocline. Just beyond the foot of the monocline the dip of the strata reverts back to the regional low dip to the northwest. Conspicuous on the geologic map (pl. 2) is the belt of Chester and Hale strata that occupies the structurally low ground at the foot of the monocline on both sides of the Neosho River. The structural displacement on the monocline amounts to roughly 300 feet.

Siebenthal (1908, p. 198) says, "West of the Neosho River the [anticlinal] fold expresses itself topographically in an abrupt faultlike escarpment to the south and a low upland slope to the north. East of the Neosho the anticline is cut through on either side by many short, steep hollows, and forms the greatly dissected highland known as the Seneca Hills."

A fault that is perhaps as much as 7½ miles long, if connected through poorly exposed stretches, strikes parallel to the axis of the structure near the foot of the monocline on the east side of the Neosho River. The displacement here is down on the southeast, augmenting the displacement on the monocline. A smaller fault, reversing the displacement on the monocline, breaks across the foot of the monocline southeast of Cleora. Possibly other faults may augment the monoclinal displacement, but the steep monoclinal slope is for the most part so covered by debris as to conceal largely the true structure.

Seneca Graben

The Seneca graben is a remarkably linear and persistent crustal break that trends northeast across the Wyandotte quadrangle from its southwest corner to the

east boundary, northeast of Racine, a distance of about 39 miles. To the southwest, an additional 22 miles of the graben extends to the vicinity of Pryor in Mayes County, and one of the bounding faults continues 4 miles farther (Miser, 1954). To the northeast the graben extends several miles beyond the quadrangle boundary, past Spurgeon. Its strike averages N. 45° E. between the southwest corner of the quadrangle and the mouth of the Elk River, N. 40° E. for several miles southwest of Seneca, and N. 48° E. between Seneca and Racine.

The graben is a complex feature in which the bounding faults are not continuous but are in many places replaced, first on one side and then on the other, or on both, by sharply dipping strata. Thus, the graben grades locally to a narrow synclinal sag. Even where the faults are present, the strata for some distance back in the adjacent blocks commonly sag toward the graben. In places the downfaulting along the graben is distributed among several parallel faults of small individual throw, some of which reverse the normal sag toward the structurally low point near the center (fig. 14). For the most part, however, the strata within the graben block are surprisingly unbroken. In many places they are virtually flat lying, but elsewhere they show dips as high as 40°, partly as a result of rapidly varying displacement along the strike of the bounding faults. The graben block is not more than 100 feet wide in some places, but is as much as half a mile wide at the mouth of Spring Branch northwest of Grove. Perhaps an average width is from 400 to 1,000 feet. Displacement from the adjacent rim to the center of the graben may be small, but commonly attains 100 feet or more, the maximum being about 150 feet. In at least one place the whole central block is raised relative to the two sides and becomes a horst.

Near the mouth of Horse Creek, at Cedar Bluffs, where the graben block is about 100 feet wide, Hale sandstone is dropped against the Joplin Member of the Boone, and the displacement is estimated at 150 feet. Two miles to the southwest, on the right bank of the Neosho River where the structural block is about 800 feet wide, the Chattanooga Shale inside the block is raised against the Reeds Spring Member of the Boone. Here the structure is a horst that has a displacement of about 63 feet on the northwest bounding fault. One and a half miles farther southwest, on the left bank of the river, where the block is about 1,000 feet wide, the structure is again a graben. Here, on the southeast bounding fault, the top of the Boone is dropped against the Grand Falls Chert Member, 15 feet below the base of the Joplin Member of the Boone. The displacement at the fault amounts to 120 feet, but because of dip in the beds of the graben block, it increases to 150 feet a short distance toward the center of the block. This locality is of particular interest in that the strata in the graben block show a reversal of dip into the fault as the fault is approached, opposite to what should be expected in response to drag (fig. 15). Possibly this reversal is related to a small longitudinal fault, lying a short distance within the graben block, that rises the center of the block relative to the narrow segment along its southeast margin. If there were late readjustments within the graben block whereby the center was pushed up slightly, this would explain both the displacement on the small longitudinal fault and the anamolous dip into the main bounding fault.

The northwest bounding fault was formerly well exposed on the right bank of the Neosho River 2 miles southwest of Cedar Bluffs, where the block is a horst. A few feet above the bed of the river the St. Joe Limestone Member inside the structural block is thrown upward against the Reeds Spring Member on the north-

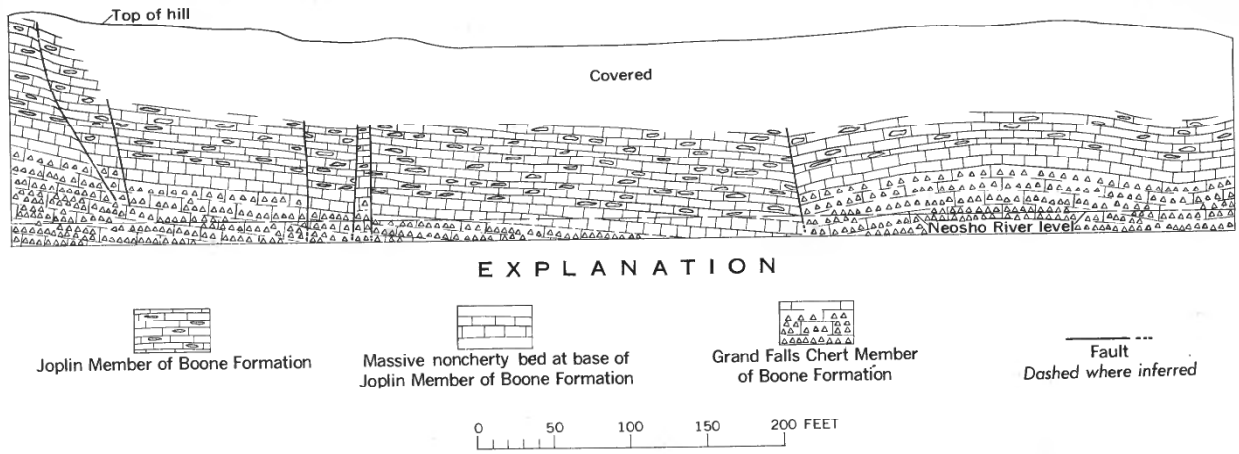

FIGURE 14.—Diagrammatic sketch showing fault pattern across the Seneca graben block opposite the mouth of Elk River.

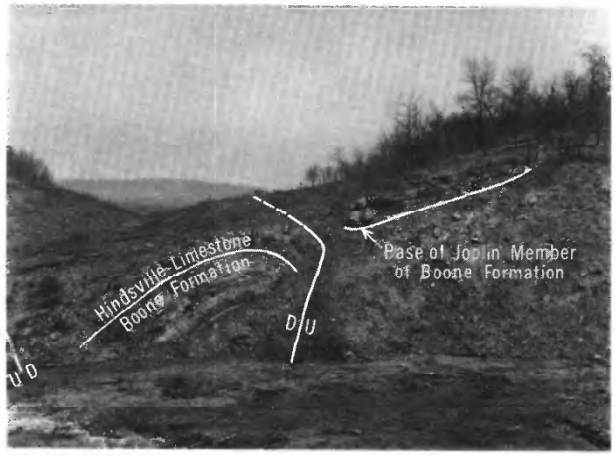

FIGURE 15.—Southeast fault of Seneca graben block, looking northeast across valley of small stream in NW¼ sec. 26, T. 24 N., R. 22 E. Photograph before impoundment of Lake O' the Cherokees.

west. Here, the fault dips about 70° NW., is 3–6 inches wide, and consists of a tight zone of crushed limestone and chert with slickensides on each side. The beds on the hanging-wall side are dragged up 2 feet in a distance of about 10 feet, and those in the footwall are dragged down 10–15 feet in a distance of 15–20 feet. Higher in the section the fault steepens to 80°, and the footwall drag is somewhat less.

All the localities just described along the Neosho River are now submerged below the surface of Lake O' the Cherokees.

Northeast of the Neosho River the graben crops out chiefly in or near the valley floors of Sycamore Creek and upper Lost Creek. Here exposures are poorer and more isolated. Most commonly at the surface, Chester strata within the graben are thrown against Boone strata outside it, but locally, the Pennsylvanian shales or sandstones are involved in the graben. As Boone strata extend to the tops of the upland in the adjacent blocks, displacements of 100 feet or more are indicated. In places the only information available as to what lies in the graben is from the records of old mine workings or churn-drill holes. Because the poor exposures in the graben block rarely permit establishing the stratigraphic sequence, it is difficult to tell Hindsville from the limestone phases of the Batesville, or Pennsylvanian black shale from Fayetteville Shale. Consequently, the strata within the graben block in this stretch have not been completely mapped. The bounding faults of the graben in this area are breccia zones that in places are as much as 20 feet wide. They may be indurated and thus form reefs on the erosion surface.

MIAMI TROUGH

The Miami trough is a linear combination of syncline and graben, similar to the Seneca graben, except that synclinal sag, with or without accompanying faults, prevails over true graben block faulting. The trough crosses the western part of the Picher mining field with an average bearing of N. 26° E. in the 8 miles or so that it traverses the field (pl. 1). Pierce and Courtier (1938, pl. 5) have mapped it an additional 15 miles northeastward to a point beyond Crestline, Kans., the strike averaging N. 38° E. for this stretch. Weidman (1932, pl. 1) shows it extending southwest from the mining field an additional 20 miles to a point beyond Afton, the strike averaging N. 22° E. for most of this stretch (pl. 2); and he (1932, p. 35) indicates that it extends several miles farther, though unmapped. The synclinal structure is well marked west of Afton, but between there and the mining field, Weidman states (1932, p. 35) that it "is marked only by slight but distinct synclinal folding as indicated in the distribution and dip of the Little Cabin member [=Warner Sandstone Member of McAlester Shale] of the Cherokee."

In the mining field the width and amount of displacement in the trough varies considerably within short distances (pls. 6–9). The width ranges from 300 feet at one point on the Anna Beaver tract (pl. 7) to nearly 2,000 feet across the Crystal and Central tracts, and averages perhaps 1,000 feet. The vertical displacement reaches a maximum on the Blue Goose No. 1 of nearly 300 feet, 200 feet of which takes place in the graben. At the level of the ore zones, the faults that account for part of the displacement are discontinuous and en echelon; those on opposite sides of the structure may join to delimit a downdropped wedge block between them, or perhaps only one fault may be present, the opposite side of the depression being marked by a steep monoclinal downwarp. Although the trough maintains a fairly constant average strike through the mining field, in detail the structure has irregular en echelon offsets between adjacent basins, with corresponding offsets between any accompanying faults that may be present. There is no noticeable displacement of the two crustal blocks on opposite sides of the trough relative to each other.

In the Blue Goose No. 1 mine where the graben phase of the structure is most pronounced, the two bounding faults dip inward toward the graben block; in the drift crossing the structurally lowest part of the block, the northwest fault dips 70° SE. and the southeast fault dips 60° NW. The latter is marked by a 2-foot-wide

zone of black to greenish shale that has been dragged into the fault zone from the adjacent Fayetteville Shale block on the downthrown side of the fault. In the Roanoke mine, where the northwest fault has the Moccasin Bend Member of the Boone on both sides, the fault dips 65°–70° SE., is narrow, tight, and has slickensides on the fault surface. At the entrance to a drift near the northwest corner of the John Beaver mine, where the fault on the northwest side of the trough lies between the Joplin and Moccasin Bend Members of the Boone, the fault is very tight and dips 55° SE. But on the southeast forty of the Gordon No. 3 tract, where the displacement on the northwest fault amounts to only 15 or 20 feet, the movement is distributed on several steeply dipping parallel breaks in a zone about 20 feet wide, and some of the intervening blocks are highly brecciated.

Bendelari Monocline

The Bendelari monocline, also known locally as the Bendelari trough (Fowler, 1938, p. 49, fig. 5), crosses the heart of the Picher field in a northwesterly direction from the Oko mine on the southeast to the Karcher mine on the northwest. It is particularly well defined on the northwest side of the Miami trough, across the Gordon tract in Oklahoma, and the Wilbur, Bendelari, Midcontinent, King Brand, and Robinson tracts in Kansas (pls. 5, 7, 8). It was named for the Bendelari mine, though it crosses only the southwest corner of that tract.

The monocline is conspicuous in the Boone and Chester strata but not in the Pennsylvanian shales. It drops the surface marking the top of the Grand Falls Chert Member down on the northeast a maximum of about 100 feet on the Midcontinent tract; or, if the displacement in a synclinal basin lying at the foot of the monocline on the King Brand tract is considered as a part of the monocline, the displacement here amounts to 140 feet (pl. 5). About 85 feet of the displacement on the Midcontinent takes place in a horizontal distance of 400 feet, with maximum dips of about 20° over part of this distance. Southeast of the Miami trough the monocline is less well defined, has irregular and gentler dips, and is arcuate in trend, from nearly east-west at the northwest end of this segment to nearly north-south at the southeast end (pls. 7, 8). As the center of the arc on the Vantage tract is in line with the monocline's course on the northwest side of the Miami trough, there is an offset of about a half a mile along the Miami trough. The displacement on the monocline southeast of the Miami trough varies widely from place to place, owing to the structural irregularity in the downdropped block. Commonly, the displacement is 40–60 feet, but locally, it may reach 90 feet.

Two pronounced structural basins, 300–500 feet across and as much as 50 feet deep, lie at the foot of the monocline on the King Brand and Bendelari tracts (pl. 5). Several shallower ones of varying size and shape lie in the same relative position elsewhere along the foot of the monocline. There is a relatively low rise of the strata to the northeast, away from the foot of the monocline, but it is so irregular that it hardly shows the synclinal structure on this limb. It is preferable, therefore, to emphasize in the name the virtually monoclinal nature of the structure as developed in the Boone strata. Chester strata underlying the basal Pennsylvanian unconformity are much thicker on the downdropped side of the monocline, showing that this structure was developed after the Chester was deposited but before or during the erosion interval preceding the deposition of the Pennsylvanian. The slight synclinal component of the structure, on the other hand, shows in the base of the Pennsylvanian shales and becomes more pronounced farther to the northwest beyond the limits of the mining field (Pierce and Courtier, 1938, pl. 5). The syncline has a low plunge to the northwest.

Rialto Basin

An irregular structural basin, nearly a mile long and as much as a quarter of a mile wide, runs in a general easterly direction through the south forty of the Rialto mine and adjacent mines to the east and west (pl. 10). Its irregularity is expressed in the occurrence of transverse structural ridges or spurs of differing height that break it into several subunits, in the lack or alignment in the several resultant subbasins, and in the presence of tributary synclinal arms that branch off at angles from the main alignment. Its maximum depth is about 80 feet at one place on the Rialto. Dips into this basin from the north are as much as 20°.

Although the basin is a prominent feature at the top of the Grand Falls Chert Member, which is the datum horizon contoured on plate 10, it does not show at the higher stratigraphic level at the base of the Krebs. The basin is contained within the north flank of a low post-Krebs dome, whose apex is indicated by outcrops of Hindsville and Batesville along Tar Creek and its east tributary a short distance to the south (pl. 2).

Smaller Synclinal Basins

Plates 5–10 show by structure contours the approximate configuration of the top of the Grand Falls Chert Member in the heart of the mining field. This horizon is by no means universally exposed in the mines. How-

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ever, the most widely mineralized unit in the district is the Joplin Member just above the Grand Falls; hence, in many stopes the contoured horizon is exposed near or at the floor of the workings. In many other places, however, particularly where the Joplin is thick, the floor may be above the Grand Falls Chert Member. Generally, the top of the Joplin Member is exposed in such places, and to show the structure on a common datum, it is necessary to transpose structural features from the top to the base of the Joplin through use of isopach or convergence information on the thickness of the Joplin Member, which can be obtained at a few contiguous underground exposures or from study of the prospect drilling logs. As the Joplin has in many places been profoundly modified and condensed by underground solution, minor warps and wrinkles that show at its top are not necessarily reflected at its base. The details shown on these plates represent a compromise, in which the structure contours are dashed where any considerable transposition of structure from the top of the unit has been necessary. Where the best, or only, structural information available is at still higher levels, from workings in the Baxter Springs or Moccasin Bend units, the dashed contour lines have to be correspondingly more generalized. The contouring in ground where there are no workings or where the workings have been inaccessible is based on logs of prospect drill holes, and is usually more generalized than where underground workings are accessible. In places, however, the drilling has been close enough and the logging accurate enough to warrant solid contours, though structural features of small size may have been missed in such areas.

Plates 5–10 show, in addition to the major features described, a plexus of small flat synclinal basins and anticlinal domes and ridges, commonly curved, in random orientation and of uncertain origin. Much of the detail shown may be the result of initial depositional irregularities, modified by compaction and settling during diagenesis.

In addition, numerous sharply incised synclinal basins are shown, which are commonly oblong in plan though nearly circular in places. These basins are as much as 1,000 feet in longest diameter, and they have vertical maximum displacements of 75 feet. Typical larger examples are on the Federal Jarrett, King Brand, Barr, Black Hawk, and New York tracts. Smaller examples on the Tom L and Admiralty tracts may be 200 feet in diameter and have a vertical relief of 30 feet. These are all structural phenomena that can in no way be explained as original depositional features.

Slump Pipes

In some of the mines of the Picher field, particularly, though not exclusively, those within half a mile of the Miami trough, small downfaulted blocks or pipes, irregular or nearly circular in plan, are of common occurrence. Some of these are shown on plates 5–7, 9. The diameter is commonly 100–300 feet, the smaller sizes prevailing. Usually, some horizon in the Baxter Springs or Moccasin Bend Member of the Boone is dropped to the mine level in the Joplin Member. The bounding fault averages near vertical but varies in detail; locally, it may incline outward either upward or downward, with dips as low as 60°. The fault is a poorly defined breccia zone that grades into the more or less brecciated edges of the adjacent blocks. It contains massive pyrite in places, as in the example along the boundary line between the Anna Beaver and Velie Lion tracts. Slickensides are rare and nonpersistent, as though related to very minor movement during later readjustments. Where less overall displacement is involved, the fault does not completely ring the pipe and is replaced by a sharp monoclinal dip of the strata on part of the periphery, as in examples on the Barr, Gordon, and Ritz mine tracts (pls. 6, 7, 9).

The block of ground inside the pipe is generally tilted, squeezed, and mashed; chert is shattered to a variable but commonly minor degree; and clay selvages have developed along many bedding seams from solution of limestone beds. True brecciation as contrasted with in-place shattering is not usually involved except at the edges. The block was generally less pervious to penetration by later underground solutions than the surrounding ore beds, hence, it remained as a barren island in mineralized ground, though invaded to a minor extent by gangue materials, such as dolomite or silica, or rarely, by sparse sphalerite.

Chester strata are included within the slump pipes of the mining field, but no clear-cut example has been observed in which black Pennsylvanian shale that had been previously deposited in normal sedimentary environment is involved. There are innumerable examples in which the shale follows an irregular vertical or diagonal course downward across the Moccasin Bend and Baxter Springs Members to connect at mine level with flat irregular underground caves into which the shale came as a sedimentary mud, carrying contemporary plant materials. The shale in the feeding pipes of such caves may show small-scale breaking and slickensiding related to minor movements during compaction. It may be difficult to distinguish such shale filling from that in which a block that formed by normal sedimentation

at a higher level has later slumped. However, the pipes filled by depositional processes are, strictly speaking, not "slump" pipes.

The geologic map (pl. 2), shows numerous small rounded patches of the Pennsylvanian, Batesville, Hindsville, and locally, Fayetteville let down into the flat upland surface underlain by the upper part of the Boone in the area east of Neosho River, particularly in the northeast corner of the Wyandotte quadrangle but also in the area bordering Buffalo Creek. Many of these occurrences are at the top of the Boone; others along shallow drainage lines are below the top. The probability is great that some of these are slump pipes, simlar to those in the mining field, though others may be the surface expression of synclinal basins. In contrast to the mining field, the Pennsylvanian shale is involved in this area. At the old Gallemore mine in the Seneca graben block, half a mile west of Racine, one of the old shafts penetrated a Pennsylvanian shale pipe 40 feet deep, that contained an 18-inch seam of coal near its base, whereas closely adjacent shafts were entirely in limestone and chert (Siebenthal, 1908, p. 204). As the coal was undoubtedly deposited in normal sedimentary environment in the Krebs Group shales, this slump pipe evidently formed some time after the beginning of Pennsylvanian deposition.

A slump pipe that dropped Hindsville and Moccasin Bend strata into the Grand Falls Chert Member was formerly exposed near the mouth of Horse Creek, about a mile northwest of the Seneca graben. This pipe was developed on the upthrown side of a small fault, showing a displacement of 50 feet, that may be related to the Seneca crustal break. The pipe is perhaps 600 feet across in shortest diameter, and is thus somewhat larger than most of those observed in the mining field.

Small Faults

In addition to faults previously discussed in relation to the Horse Creek anticline, Seneca graben, Miami trough, and slump pipes, several faults have been revealed by C. C. Branson's detailed mapping of the Pennsylvanian outcrops and some contiguous terrain between Narcissa, Fairland, and Afton. Most of these faults are short and of small throw, though two are as much as 3 miles long. They trend mostly northeast, parallel, or at a low angle, to the strike of the Miami trough. Two faults northeast of Narcissa, however, trend northwest and offset the contact between the Pennsylvanian and older strata (Reed and others, 1955, pl. 1).

The random small faults found by detailed underground mapping in the mining field are shown on plates 5–10. Many of them are obviously related to slumpage of the same general type that produced the slump pipes, as shown where a linear or only slightly curved short fault borders one side of a small and abruptly downsagged block. Other linear faults have no set pattern of orientation. None of them, except where related to the Miami trough system, are more than 1,000 feet long. They have maximum displacements of 20 feet, though generally only 5–10 feet. Owing to the structural incompetence of several limestones overlying the Grand Falls Chert Member, particularly the Joplin Member, probably few of the small random faults breaking the top of the Grand Falls persist upward even as far as the top of the Boone, though others may develop in place of them at higher levels. Some of the faults shown on plates 5–10 by dashed pattern have been projected from higher levels and are subject to comparable uncertainty.

A type of fault not shown on plates 5–10, because it does not cut the Grand Falls, is low angle, of only a few feet displacement, and as commonly of thrust displacement as otherwise. Such faults are confined to the incompetent strata and are not commonly recognized in the generally brecciated terrain unless they displace some easily recognizable stratigraphic marker, such as J or K bed.

Slump Breccias

Akin to the slump pipes are the slump breccias formed in strata where there has been much solution and condensation, as in the Chester strata on the upper level of the Crystal and Central mines. The matrix is a residual gray selvage containing ill-defined large blocks of Chester limestone, streaks of green clay believed to be residual from solution of limestone blocks, and more angular fragments of green shale, quartzitic sandstone, and black shale, the latter derived from the overlying Fayetteville Shale before it was eroded at the pre-Pennsylvanian unconformity. These slump breccias are presumably the basal parts of slump pipes that originally extended into the overlying Chester strata. Similar breccias in which chert nodules and fragments are contained in a clay residual from the solution of the limestone are found particularly in the Joplin Member of the Boone but also in other Boone horizons.

Tectonic Breccias

The strata exposed in the walls of the mine workings have been brecciated to a varying degree. Subsequent mineralization had practically no effect on the chert in the original breccias, but it so modified the matrix

through solution or replacement of the original carbonate rock by gangue and ore materials as to mask many features of the initial brecciation. However, not all the breccias were mineralized; and although such ground, being barren, is not commonly cut in the mine workings, there are local exposures in developmental workings, or less commonly on the edges of stopes, that reveal the unaltered breccias. The mineralized breccias which are more abundantly exposed can be interpreted from the evidence shown by these few barren exposures.

The exposures in unaltered ground show breccias in which the angular chert fragments are enclosed in a compact limestone matrix. Where the limestone was originally medium to coarse grained, as in the Joplin Member (M bed) or in K bed, it was recrystallized, though incompletely, and any bedding features, obscure at best in the original, were destroyed. The enclosed chert, on the other hand, whether in unaltered or mineralized breccias, may in places retain a rough layering in which the brecciated fragments of recognizable beds or nodule zones, though randomly rotated, keep the same approximate stratigraphic position as in less disturbed sections. Commonly, adjacent chert fragments can be recognized as parts of an initial nodule or layer that, with some rotation of the parts, can be restored. With only slightly more intense deformation, any semblance of layering is lost, even though individual chert fragments in the breccia masses may not be greatly displaced from their original position.

The brecciation undoubtedly was produced by deformation of the chert-bearing limestone in which the brittle chert was brecciated while the enclosing limestone behaved plastically. Differential horizontal slippage along bedding directions brought about by even the slight structural warping to which the region has been subjected would be concentrated in the less competent strata. The Joplin member, lying between the Grand Falls Chert Member and L bed chert or the lower part of the Moccasin Bend cherts where L bed is absent, would be particularly vulnerable to such deformation; and K bed, lying between L bed and the lower part of the Moccasin Bend, would be equally vulnerable.

That such stresses have been effective in the ground of the mining field is indicated in the Kansouri mine just west of the Miami trough, where at one place a large block of chert from the lower thin bedded part of L bed has been plucked out and pulled laterally into the underlying breccia beneath an unbroken L bed roof. The low-angle normal and thrust faults of small displacement found in the incompetent strata at many places are a manifestation of the same type of stresses.

Joint Systems

The structural feature that is most closely related to ore localization is the fracturing to which the host rock has been subjected along joint systems. The fracturing is best defined in the thick chert strata, particularly in the L bed roofs of stopes in the Joplin Member. Most of the fracturing averages near vertical, though with many deviations in individual fractures and numerous cross or diagonal breaks. In some fracture systems, the fractures have a consistent dip to one side. The trace of the fractures on the relatively flat plane of the mine roof is a series of anastomosing subparallel lines, in zones that show an elongation parallel to the strike of the fracturing. The zones are not constant in strike, and commonly curve along the strike. These fracture zones commonly determine the trends of the ore runs, though the two may also appear to be independent of each other. The fractures are partly disguised in the ore-bearing strata because of less regular breakage of the rock and modification by solution effects accompanying ore deposition. They were probably developed and intensified over several different periods, contemporaneously with the tectonic breccias.

These fracture zones have been grouped by Fowler and Lyden (1932, p. 225–229, fig. 8) with the faults along the Miami trough under the term "shear zones." However, they generally show no displacement except for uncommon slumps of a few inches that may be in part related to leaching and slumpage in underlying strata. Some are related to minor warps in the bedrock whereas others (Netta mine) are present in surprisingly little-deformed ground. They may die out abruptly along strike, and have no predictable continuity. Few of the fractures are accompanied by inconspicuous slickensides in the chert. The differential movement involved in their formation was negligible compared to that usually implied in the term "shear zone." They are here called joint systems, for they most resemble the joints found in various types of massive rocks. Most of these joint systems were earlier than the ore and served as channelways for introduction of the ore-bearing solutions, but some, of a different origin, were formed contemporaneously with the mineralization.

Origin and Age of Structural Features

In the Wyandotte quadrangle the conspicuous linear structural features—the Horse Creek anticline, Seneca graben, Miami trough, and Bendelari monocline—are obviously of tectonic origin. The first three of these features coincide in trend with the northeasterly alinement of faults and synclines that show conspicuously

in the northeastern part of the geologic map of Oklahoma (Miser, 1954). The Miami trough and Seneca graben may have been modified and accentuated locally by the collapse of beds due to solution of subsurface carbonate rocks, as suggested by Quinn (1963) for similar structural features in northwestern Arkansas. This cannot be the sole explanation for the graben displacement, however, for the Miami trough appears to be well developed in the igneous rocks of the Precambrian basement at depth (fig. 3). More pronounced displacement in the sedimentary strata at higher levels in the Miami trough (fig. 3) may indicate the extent of accentuation by solution. Some of the smaller structural features, especially some of the slump pipes, are probably wholly of solution-collapse origin, though of course tectonic jointing may have had some influence in guiding and localizing the flow of the dissolving ground waters. Other structural features of intermediate size and inconspicuous linear trend, such as some of the larger synclinal basins in the Picher mining field, may have a mixed tectonic and solution origin, or are indeterminate.

Most of the tectonic structures involve beds of the Krebs Group and so are post-Krebs in age, at least in part, but the earliest deformation of some of these structures occurred sometime after deposition of the Chester Series and before deposition of the Krebs. Much of the solution and collapse of the slump pipes and related structures occurred after Chester time and before any or much of the Krebs was deposited. Since the Krebs accumulated, both ordinary ground waters and the ore-bearing waters have caused some solution.

The slump pipes resulted from gravity caving into leached ground. As many of the slump pipes observed in the mines extend into the Grand Falls Chert Member, obviously some bed in or below this unit was among those leached. Prospect drilling has not been deep enough to shed any light on this question, but one may conjecture that the limestones in the Reeds Spring and St. Joe may have been particularly vulnerable. Dolomites in the Cotter, though less soluble, may also have been dissolved in some places. Fracturing of these units and of the Grand Falls Chert Member during earliest structural deformation would, upon solution of the supporting carbonate rock, have left the chert phase weak enough to collapse under the superincumbent weight.

In other places the slump pipes are bottomed at higher stratigraphic levels. The Joplin Member, in particular, is commonly leached of all carbonate material, and its contained chert is concentrated in a residual clay matrix. Such ground may form the bases of pipes that involve Chester strata at the top, or shale pipes in which the Pennsylvanian shale came into the openings as an original clay sediment at the time of initial Krebs deposition. Commonly, a few feet of dark shaly material, probably in most places of mixed sedimentary and residual clay derivation, shows in many of the mine workings at the base of the Joplin Member, particularly at the zone of chert-free limestone (p. 36).

Bretz (1950) has made a thorough study of filled sink structures in the Ozark upland in Missouri, in which near-cylindrical masses of deformed Pennsylvanian strata have slumped down into sinkholes in underlying Ordovician and locally Mississippian carbonate rocks. He believes that such features were formed by slow subsidence under load, coincident with solution of the carbonate rocks well below the ground-water table, and without formation of an underground cave at any time. This explanation, by and large, is valid for the slump pipes in the Wyandotte quadrangle. There is, however, evidence in the contemporary shale fillings that some low flat caves existed at the beginning of Pennsylvanian sedimentation; and it is reasonable to assume that the collapse of such caves may have initiated subsidence in many places. After the slump boundaries had been defined by the initial breakage, the core block would continue to slump gradually as carbonate material was further leached in the soluble zone. All limestone-bearing strata at various horizons in the mines, but particularly on the K and M bed levels, contain residual seams of gray to greenish clay selvage, in part interstitial to the chert nodules and fragments, obviously residual from the solution of limestone largely in place, though some of it may have been transported a short way. This selvage grades laterally without definite boundaries into the unleached rock, and probably is a record of solution under the pressure of the superincumbent strata. The selvage in K bed generally contains glauconite which is residual from the limestone.

Slump breccias in which rocks from different stratigraphic levels are mixed may well be a record of initial small cave fillings. They are especially conspicuous in the Chester workings of the Crystal and Central mines, where they commonly include angular fragments of hard black shales, presumably from the Fayetteville, in addition to fragments and residual clays from other Chester strata.

The slump-pipe concept can be modified and extended to the smaller synclinal basins found in the mining field, such as those on the Tom L and Admiralty (pls. 9, 10). In these examples the Boone strata have sagged into the leached area rather than breaking around the margin. Larger basins, such as those on the Federal Jarrett, King Brand, Barr, Black Hawk, and New York, and particularly the Rialto basin, present more of a prob-

lem. The structural relief in these basins of as much as 80 feet requires solution of carbonate rock well below the base of the Boone, for the amount of limestone in the Reeds Spring and St. Joe Members, even if all dissolved out, is not great enough to account for this much slump. Furthermore, the cuttings examined from drill holes in the Federal Jarrett and Barr basins contain unleached limestone in the Reeds Spring at the greatest depths reached, 340 feet below the base of the Pennsylvanian shale. Nevertheless, the lack of any definite hiatus in structural type between the smallest and largest basins points to a common origin. A feature as large as the Rialto basin, which shows a definite elongation, is probably localized on a deep fracture zone along which the Ordovician dolomite has been leached. Other more equidimensional slump basins may be localized on deep intersections of fractures. All these slump basins differ little from the series of basins which, in the aggregate, constitute the Miami trough, except that associated faulting has not generally been demonstrated. However, the absence of faulting may be more apparent than real, for the scattered basins commonly are in unmineralized ground and thus lack underground exposures adequate for complete structural delineation.

The structural deformation, including the development of slump features, may have ranged over a considerable stretch of geologic time, but it can be dated only in general terms relative to a few fixed stratigraphic tie points of unequal and indefinite time spacing. Except for certain alluvial and terrace deposits of Quaternary age, the youngest strata in the Wyandotte quadrangle belong to the Krebs Group of Des Moines (Pennsylvanian) age. Most of the structural features of tectonic origin involve the Krebs and are thus, in part at least, of post-Krebs age, but some of these features were in existence before the beginning of Krebs time and were simply accentuated to varying degrees in post-Krebs time.

Along the Miami trough the Bluejacket Sandstone Member of the Boggy Formation has been folded or faulted down to the level of lower shales in the Krebs Group at some time after the Krebs was deposited; but there had been an earlier synclinal or graben depression along the axis of the trough at some time between Chester and Krebs deposition, for the Chester strata, lying unconformably beneath the Krebs, are generally thicker in the trough than in bordering areas, and the Fayetteville Shale is preserved only in the trough. The Bendelari monocline involved the Chester and older strata but was largely of pre-Krebs age, as accentuation of the structure is post-Krebs time amounted to only a relatively slight synclinal warping. Pennsylvanian shale occurs commonly within the Seneca graben block northeast of the Neosho River, indicating that this feature is post-Krebs. If it was initiated in pre-Krebs time, the evidence for this has not been observed, at least in the Wyandotte quadrangle. The Horse Creek anticline is post-Hale and probably post-Krebs in age, but as the black shales of the Krebs Group have been eroded on the axis and south side of the anticline, relation of the structure to the Krebs cannot be confirmed.

The Rialto basin is developed in Boone and Chester strata but does not show at the base of the Krebs. If the Quapaw Limestone is present in the basin (p. 56), warping and truncation of this unit before the Hindsville Limestone was deposited is a possibility; but uncertainty as to stratigraphic and structural interpretation from the available churn-drill cuttings makes this highly conjectural. Other smaller basins that show on plates 5–10 are largely post-Batesville and pre-Krebs, to judge from the abrupt thickening of the Chester strata in the basins beneath a fairly level shale cap. In a few there is a slight slump suggested at the base of the shale, but generally the drilling is not closely enough spaced to define very accurately the details of configuration at this horizon. In any event, what post-Krebs accentuation of the basins that may exist is minor in comparison to the pre-Krebs structure.

The slump pipes are too small to yield much information pertinent to their age through prospect drilling. They are known chiefly from exposures in the mine workings. Most of the slump pipes in the mining field are believed to have formed after Chester deposition and prior to the accumulation of strata above the basal 100–200 feet of Pennsylvanian strata, probably forming at different times throughout this interval. One slump pipe whose lower part is exposed in the Gordon No. 3 workings was cut by a drill hole that showed 40 feet of Hindsville Limestone in comparison to 5–8 feet of this unit beneath the Pennsylvanian shales in drill holes 200 feet away but outside the slump. This suggests a post-Chester and pre-Krebs age for this slump pipe. The evidence for dating others is indirect and related to their origin, as discussed in the next few paragraphs. The closely related slump breccias in Chester strata of the Crystal and Central mines were probably formed during the same time interval as the slump pipes.

The extensive underground leaching that produced the slump pipes and slump breccias was initiated after the post-Chester uplift. Presumably, it did not become effective until the Fayetteville Shale had been at least partially removed to expose the underlying limestones. Underground solution continued throughout most of

the erosion interval preceding the deposition of the Krebs shales. It was most effective in the early stages when there was some topographic relief to produce a hydrostatic head in the water table, leading to an underground circulation, for the absence of any known surface drainage channels below what ultimately became the basal Krebs peneplane indicates that the solution openings were leached below the water table. The ground-water circulation slowed up as planation progressed, though there was still enough movement of ground water to carry in and deposit the initial muds at the beginning of the Krebs sedimentation. All water circulation presumably ceased as soon as a sealing layer of Pennsylvanian shale was deposited over the peneplaned surface. Caves not previously filled collapsed as soon as enough shale was deposited over the surface to crush in the roof, and in this process, previously deposited shale would take part in the slump, at the top of a slump pipe.

The tendency for the slump pipes that have been recognized at the M bed mining level to be concentrated in the vicinity of the Miami trough suggests that the structural breaking during the earlier (pre-Krebs) deformation along the trough made this ground especially susceptible to solution.

In areas still covered by the Krebs shales, such as most of the Picher mining field, the existing slump pipes can be assigned to the erosion interval that ended with the deposition of these shales, even though some of the slumps may have broken down slightly later. In the area east of the present shale erosion boundary, however, formation of slump pipes was renewed during erosion of the Krebs shales and has probably continued to the present. It is impossible to segregate definitely the later from the earlier generation of pipes, though some generalizations as to probabilities seem warranted. Pipes containing Fayetteville Shale in the northeastern part of the Wyandotte quadrangle undoubtedly belong to the earlier generation; for in the Joplin area a short distance to the northeast, this formation was eroded during the time interval represented in the unconformity at the base of the Krebs, except for remnants of Fayetteville equivalent (Carterville) preserved in sinkholes (Smith and Siebenthal, 1907, p. 5–6). The same stratigraphic relations probably held in adjacent parts of the Wyandotte quadrangle, but the evidence has been removed by erosion. Although erosional remnants of the Batesville and Hindsville remain in the quadrangle and could have furnished the filling for slump pipes during the post-Pennsylvanian period of erosion, these formations occur only in pre-Krebs sinkholes in the Joplin area, and probabilities seem to favor this dating for at least most of those in the Wyandotte quadrangle. Pipes containing Krebs Group shale in places record contemporary filling of openings at the beginning of shale deposition, or slumpage during the early stages of this deposition. On the other hand, particularly in areas where the stratigraphic position of the shale was close to the Boone, shale pipes have been dropped down into the Boone in the later cycle, during erosion of the shale.

Features similar to the slump pipes in Cherokee County, Kans., just north of the mining field, have been described by Pierce and Courtier (1938, p. 53–54, 58–60) as sinkholes; some involving the Krebs shales have formed in historic time, and other "ancient" ones are considered by Pierce and Courtier to be post-Krebs because in places they involve strata above the middle of the Cherokee (Krebs plus Cabaniss Groups). The examples in the Missouri Ozarks in which Pennsylvanian strata have slumped into Mississippian or Ordovician bedrock, as described by Bretz (1950), must have formed during this post-Krebs erosion cycle. At the Simpson coal mine in Moniteau County, Mo., Pennsylvanian strata filling such a sink have been dropped as much as 400 feet below their proper stratigraphic level (Bertz, 1950, p. 804). The slump pipe near the mouth of Horse Creek could also be of much later date than those in the mining district. The post-Carboniferous erosion history of the Ozarks is not well-enough known to date very closely initiation of the post-Krebs erosion cycle.

The origin of the slump pipes given in this report differs, at least in relative emphasis of processes, from that given by Smith and Siebenthal (1907) for the Joplin area. These authors stress the contemporary filling of caves and sinkholes in karst topography by sedimentation during initial invasion of seas. Thus, they believe that the Carterville, equivalent to Fayetteville, Batesville, and probably Hindsville, was deposited on a karst topography developed on the Boone surface in the pre-Chester erosion interval, and that it was later eroded, except for cave-filling roots, during the pre-Krebs erosion interval. The study of drill cuttings from many prospect holes in the Picher mining field reveals no evidence for local erosional relief between the Boone and Chester strata, but suggests a low-angle and gradual eastward truncation of the uppermost part of the Boone beneath the basal Hindsville. With the one exception in the Mary Jane mine, contemporary filling of caves in the Boone by sedimentary material of Chester age has not been found. There is, however, abundant evidence of cave formation in the post-Chester and pre-Krebs erosion interval; and the collapse of slump pipes above such openings, followed by further slumpage

during solution at the base, forms a ready mechanism for dragging Chester strata down into the Boone. Siebenthal's later mapping of Hindsville, Batesville, and Fayetteville fillings in different sinkholes in the Wyandotte quadrangle east of the Spring and Neosho Rivers (pl. 2) favors the slump-pipe origin, for otherwise it would seem that all should be filled by the Chester formation that initially invaded the supposed karst topography, namely, the Hindsville.

Smith and Siebenthal were correct in recognizing the contemporary invasion of numerous sinkholes and caves in the Boone by black muds at the beginning of the Krebs deposition, but this process can account for only certain of the sinkhole fillings. The occurrence of bedded coal in some, as at the Gallemore mine near Racine (Siebenthal, 1908, p. 204), in thicknesses comparable to that found in undisturbed sedimentary sequences in bordering areas, is an indication that the coal formed in a normal environment and slumped later.

Breccias formed by differential movement along bedding during deformation have in preceding pages been called tectonic breccias. They can be dated only by relating them to larger structural features that are otherwise dated. The Miami trough, which lies in the middle of an area containing abundant tectonic breccias, was the site of deformation in the period between deposition of the Fayetteville (Chester) and of the Krebs, and was the site of further deformation at one or more periods in post-Krebs time. The Bendelari monocline was formed mostly in the earliest period of deformation indicated, but was slightly modified at the later period or periods. It would, therefore, seem evident that at least some of the breccias were formed before the Pennsylvanian strata were deposited, though they doubtless were augmented in post-Krebs time. No example has been observed in which a slump pipe, believed to be not later than early Krebs in age in the mining field, has been offset or otherwise disturbed by differential movement along the bedding of the type that accounts for the tectonic breccias. However, critical exposures of the slump pipes are not extensive enough, and their interpretation is too uncertain to warrant broad conclusions from such negative evidence. Rare and local brecciation between successive stages of the ore mineralization shows that some differential movement persisted into the period of mineralization, which is known to be post-Krebs. The relative ages at intersections of tectonic breccias with slump pipes needs further investigation.

Most of the small-displacement faults and the joint systems that control many of the ore runs were probably produced by the same stresses that produced the tectonic breccias and larger faults. Breakage of the rocks was cumulative, though probably the final stages, when the rock had acquired more brittleness, were more effective in preparing the ground for mineralization. Some of the joint systems in the roofs of stopes were produced by slumpage in the ore-bearing and adjacent ground owing to leaching of carbonate rock during the period of mineralization. Criteria for distinguishing these joints from the preore joint systems of tectonic origin are dependent on recognition of condensation, but are indefinite and uncertain for many specific examples.

The structural history within the Picher mining field may be summarized as follows: Aside from the gentle structural tiltings that produced several unconformities before Chester time, the first period of appreciable deformation recognizable in the exposed rocks was between the deposition of the Fayetteville Shale and the Krebs Group. It was accompanied by plastic deformation of the limestone and brecciation of the contained chert in noncompetent strata, particularly in the ore beds. It was also accompanied or followed by a period of erosion, peneplanation, and subsurface leaching of limesone during which slump pipes and slump breccias formed. Slumping continued only slightly into the period of Krebs deposition. At some time later, during another period or other periods of deformation the Krebs was folded and faulted in part along or near the axes of the earlier deformation. This was accompanied by further brecciation of incompetent beds. The later deformation, or a final stage of it, reopened the joint systems which in part guided and controlled the ore deposition.

There is no way of positively dating the post-Krebs deformation, nor is it possible to determine whether it is the product of one short period or the cumulative effect of several such periods. Although the type of deformation is totally different from the orogeny that produced the Ouachita Mountains, that period of folding would seem to be the most probable time for sympathetic adjustments in the Ozark block. If the Ouachita folding is correlated with that of the Arbuckle anticline, as suggested by Miser (1929, p. 27), the period of intense folding was in Late Pennsylvanian (Virgil) time (Ham, 1954, p. 2041). On the other hand, King (in Flawn and others, 1961, p. 186–188) ascribes the deformation in the Ouachita system to a long time interval between Late Mississippian and Middle Pennsylvanian (Des Moines) time. As the Krebs is of Des Moines age, correlation of the post-Krebs deformation with the Ouachita disturbance would, under this interpretation, limit it to a narrow time interval near the end of Des Moines time. However, the final adjustments could have occurred considerably later.

Since the start of ore deposition the ground has been stable and undisturbed, except for negligible fracturing in some places and slight brecciation in still fewer places during and following ore deposition. Sizable caves formed by the ore solutions and containing delicate ribs and bridges of ore and gangue minerals remained unbroken and remarkably clean until disturbed by mining operations.

ORIGIN OF CHERT

Chert is present to varying degree in all members of the Boone Formation. Its abundance, shape, and dimensions of megascopic units and its purity, luster, and color are essential elements of the stratigraphy. It may occur as beds with smooth or irregular boundaries, as rounded or elliptic and commonly banded nodules, or as irregular amoeboid masses of innumerable shapes; or more massive beds of chert may contain residual limestone in equally irregular shapes. Most occurrences that have not been structurally disturbed are concordant with the bedding, but there is a local minor occurrence in fractures that crosscut the bedding. Cotton rock, which is an intermediate product between limestone and chert, has the same forms of occurrence. It may be leached of its lime to form tripoli.

Microscopic study shows that all gradations exist between limestone and chert. In cotton rock of low silica content, the fine silica granules are aggregated in interstitial nests between the calcite grains. These nests increase in size as the percentage of silica increases toward chert until the calcite is left as small ragged grains scattered through a chert base. The pure cherts are microcrystalline; most of them are composed of minute irregular grains of quartz (Folk and Weaver, 1952) that have no recognizable optical elongation (fig. 16). The average grain size is about 0.002–0.006 mm, but grains commonly are as large as 0.015 mm. A variant of the chert texture in which the minute silica granules have elongation that is optically negative, opposite to the prismatic elongation of quartz, is present in some cotton rocks and cherts (fig. 17). Typical grain sizes here are 0.005–0.015 mm long and 0.002–0.006 mm wide. This texture is particularly characteristic of L bed cotton rock and chert, but many cotton rocks and a few cherts at other stratigraphic levels from at least the Grand Falls to the top of the Boone also have it. All gradations exist between this texture and the normal one in which the grains have no optical elongation.

Scattered larger grains of quartz, though only a fraction of a millimeter across, are present in widely varying amounts in nearly all thin sections of Boone cherts, and may aggregate into irregular clusters or wisps. Scattered single grains that are angular or have an elongation inclined to the extinction directions of the quartz are generally of detrital (silt) origin. These may have quartz overgrowths beyond the boundaries of the original fragments that are recognizable under proper lighting. They are especially noticeable in lower L bed.

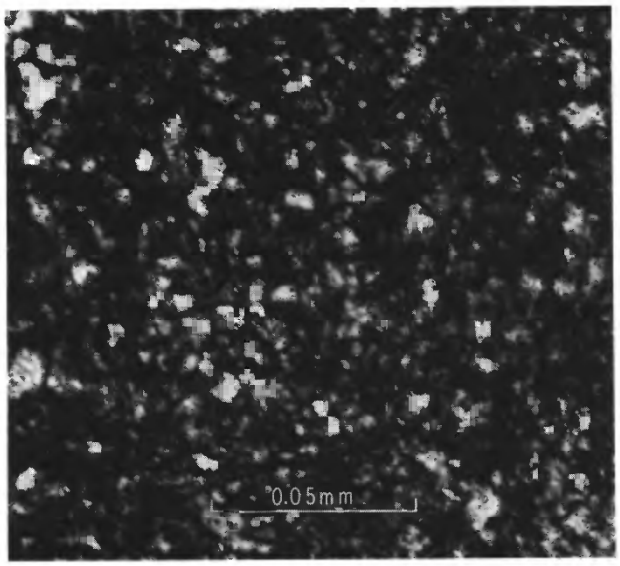

FIGURE 16.—Photomicrograph of thin section showing normal texture of chert; from basal part of Joplin Member of Boone Formation, New Chicago No. 1 mine. Crossed nicols.

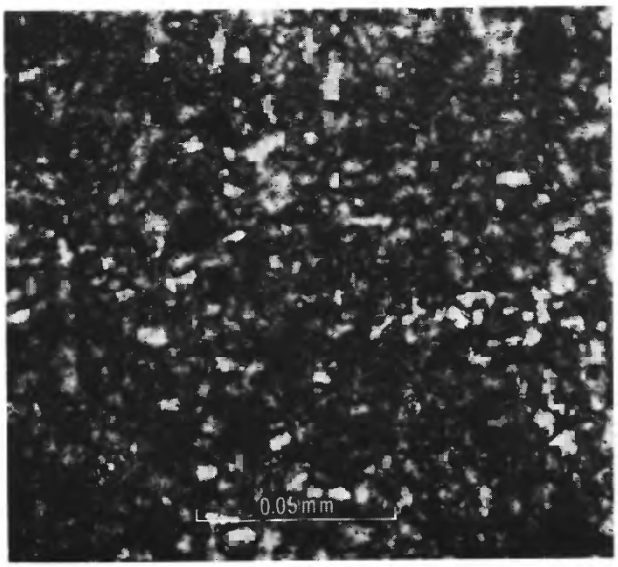

FIGURE 17.—Photomicrograph of thin section showing linear habit of chert grains in many cotton rocks and cherts; from basal L bed, Bendelari mine. Crossed nicols.

Other scattered grains that have positive optical elongation may in part represent a later interstitial growth of hydrothermal jasperoid, but some of these grains are probably of primary crystallization in the chert, as are the larger irregular grains of more equidimensional aspect, which commonly form mosaics of several grains. All gradations in size may be found between such grains and the typical fine chert grains. Many cherts also have scattered blebs or wisps of more coarsely crystalline chalcedony, some of which are probably relict from extensive replacement of organisms that are no longer recognizable. In some cherts the chalcedony predominates.

Although Folk and Weaver (1952) have emphasized the lack of any distinction between quartz and chalcedony when the X-ray diffraction patterns are compared, the petrographic distinction is a useful one in any discussion of chert. Chalcedony is fibrous and has a tendency to radial or parallel growth; its fibers have negative optical elongation compared to quartz; and the index of refraction is lower. Chalcedony is particularly conspicuous in the replacement of organisms in chert, though it also has other types of occurrence. More intense replacement largely destroys the organic textures and yields mosaics of quartz grains much coarser than the typical chert. The minute granules of chert that have negative elongation (fig. 17) may be chalcedony, though they lack any fibrosity, radial or parallel tendency, or other identification with chalcedony.

All the cherts, including the chalcedony phases, contain minute, and in large part submicroscopic, inclusions of some substance that imparts a white cloudiness in reflected light. Intensity of the clouding varies widely and abruptly even within a petrographic thin section. This substance appears brown in thin section owing to its low refractive index relative to quartz, with resultant dispersion effects in fine particles. The grosser particles, which can be resolved under magnification of × 500, appear in the chert grains and chalcedony masses as irregular granules, scales, twiglike inclusions, or rarely, as thin lines around circles and arcs of small diamter (0.01–0.03 mm.), probably of organic affiliation. Heavily clouded parts of the chert may be checked by fine shrinkage cracks that are filled with clear silica of higher refractive index, though the silica crystal units of the clouded mass on each side have crystallized across the cracks without optical break. The shrinkage cracks suggest that the silica mass was once in a gelatinous state before its final crystallization.

Folk and Weaver (1952) have suggested that a brown clouding (white in reflected light) in the cherts that they studied under electron microscope magnifications is due to minute globular inclusions of entrapped water. They ascribe the differences in specific gravity and refractive index between quartz and chalcedony to the predominance of these diluting inclusions in the chalcedony. However, the correlation between the brown clouding and the occurrence of chalcedony in the Boone cherts examined by us is not convincing; some heavily clouded material is, in all other respects, the microcrystalline quartz of the typical chert, whereas some of the radiating fibrous chalcedony in interstitial pockets is relatively clear. Fine inclusions of the type described here in the Boone cherts also exist in the replaced limestone. The fact that they occasionally remain entrapped, though with some agglomeration in particle size, inside hydrothermal quartz grains (jasperoid) that replaced the limestone during sulfide mineralization suggests that the substance is organic.

Sponge spicules (monaxons) are common in thin sections of cherts from at least the Grand Falls to the top of the Boone, and some have been found in closely associated limestone. The more complex spicules have a nearly isotropic core surrounded by a chalcedony shell. The core has the same index of refraction as the enclosing chert, and is apparently composed of chert (quartz) granules so fine that the birefringence is barely discernable. The core commonly has a faint olive-green or olive-brown tinge, and contains a few minute spindles of a mineral resembling a sedimentary mica. The shell, where present, usually has the chalcedony fibers arranged approximately radial, less commonly parallel to the spicule axis. Many spicules are constituted entirely of the core material. Others are chalcedony throughout, with the fibers elongated parallel to the axis of the spicule, less commonly transverse or oblique to this axis, or the orientation may be random. Different orientations of the chalcedony may show in the same spicule, with gradual or abrupt transitions. Whether the petrographic differences in spicules are primary or are due to differences in diagentic history is not evident. Spicule diameters range from 0.02 to 0.20 mm, most from 0.04 to 0.10 mm. Length as revealed in thin section has little meaning, as the spicules are rarely in the exact plane of the section; but intercepts rarely exceed 1 mm in length, though noted as long as 2 mm. Some of the spicules contain a great excess of the "brown" clouding compared with the chert matrix, others contain less, and there is commonly a difference in the clouding between shell and core.

Crinoid stem sections in the chert are commonly composed of crystalline calcite, as in the limestone; but they may be replaced by a mosaic of cryptocrystalline silica grains which in many instances are in parallel optical

orientation, so that the fossil behaves as an optical unit under the polarizing microscope. Other types of carbonate shells exhibit similar oriented replacement. Whether the replacing grains are quartz or chalcedony cannot usually be determined, though some are obviously chalcedony with the fibrosity perpendicular to the elongation of the shell. The interiors of such shells are commonly filled by clear fibrous chalcedony in radiating sheafs or by granular quartz mosaics of considerably coarser grain than the chert texture. In places, however, the fine chert forms the filling.

The Short Creek Oolite Member, where silicified, commonly shows an aggregation of fine chalcedony grains that are radially oriented in the oolite spherules. The matrix for the spherules may be chalcedony in coarser fan shaped crystal groups, or it may be the cryptocrystalline chert or coarser mosaic quartz.

The lime content in nodular chert varies between different bands from negligible in the dark bands to a high percentage in the peripheral nearly white band that is usually present and that is gradational into the enclosing limestone (Giles, 1935, p. 1828).

Most masses of chert are brittle when subjected to structural stresses, though the brittleness decreases with increasing admixtures of microscopic calcite on the peripheries of chert masses or in the cotton-rock phases. With the enclosing limestone acting as a plastic medium under deformation, many new textural relations between the limestone and chert are introduced in structurally deformed ground. Furthermore, both the limestone and to some extent the chert are soluble, the latter particularly along fragment contacts. Differential leaching of the limestone may increase the percentage of chert in a given stratigraphic unit without any actual addition of silica, and even with some leaching of the chert. In structurally disturbed and leached ground where the bedding is destroyed in the process, it is usually impossible to judge whether new chert has been added, or whether only an apparent cheritification has been accomplished by condensation.

Explanations on the origin of the chert have been varied. Tarr (1926) proposed a syngenetic origin wherein the chert was precipitated on the sea floor as a colloidal silica gel at the time of sedimentation. At the opposite extreme, Fowler and Lyden (1932) and Fowler and others (1935) postulate an epigenetic origin wherein the chert is of hydrothermal source, introduced into the limestone along structurally deformed zones subsequent to deposition of the full thickness of the Boone Formation. Because pebbles of Boone chert are locally incorporated in the basal conglomerate of the overlying Chester, and because the Chester is relatively free of chert, the age of Boone chertification is considered by them to be narrowly limited to post-Boone and pre-Chester time. The theory of Fowler and Lyden is followed by Weidman (1932, p. 84–85) and is favored by Giles (1935), who, however, points out (p. 1859–1860) certain unfavorable factors in its application.

We believe that the great bulk of the chert in the Boone was formed by replacement of successive surface or near-surface layers of limestone on the floor of the sea during the deposition of the Mississippian strata, but there was some later movement of silica and precipitation of additional chert by ground waters during cycles of erosion and terrain leaching, particularly between the end of the Mississippian and beginning of Krebs (Pennsylvanian) deposition.

PRIMARY CHERT

Evidence for replacement of the limestone by the chert has been presented by Bastin (1933, p. 377–380) and Giles (1935, p. 1843–1847). Points of evidence that appear convincing include—

1. The silification of fossils enclosed in the chert;
2. The preservation of limestone bedding features and glauconite grains in the chert;
3. Transection of the limestone stratification by the concentric banding at the ends of chert nodules;
4. The presence of irregular "islands" of limestone in sheets of chert wherein the limestone islands possess the same trend and textural features as in the limestone forming a continuation of the same bed;
5. The gradation from chert nodules or less regular chert masses into adjacent crystalline limestone;
6. The "lateral interpenetration of chert and limestone by [along?] irregular, interfingering, and frayed contacts" (Giles, 1935, p. 1845);
7. The local chertification of the Short Creek Oolite Member with preservation of the oolitic texture; and,
8. In siliceous limestone, the microscopic invasion or crosscutting of calcite grains by the chert with very irregular frayed boundaries.

In the replacement of the Short Creek Oolite Member, the chert may assume a nodular shape (fig. 8), showing that this shape is of concretionary origin and typical of replacement, rather than constituting evidence for precipitation of silica on the floor of the sea as a gelatinous mass that assumed the elliptic shape in response to the pull of gravity, as postulated by Tarr. Further evidence of replacement by concentrically banded chert is shown in K bed at the Mary Jane mine, where a residual of limestone is enclosed in chert which

shows concentric banding in the typical nodular shapes around the limestone core—a negative chert nodule in this instance (fig. 18).

That the replacement of the limestone by the chert was closely related to sedimentation is indicated by the concordant relationship between the chert and the bedding in undisturbed ground. The nodules are segregated at varying distances apart on definite stratigraphic levels, those at one level commonly showing uniformity of features, contrasting with those in adjacent levels. They are elongated with the bedding, and in rare instances where individual unbroken nodules are oblique or perpendicular to the bedding, a structural or solutional disturbance can usually be demonstrated. Less regular-shaped chert masses show similar relationships. Certain chert beds are sharply enough defined from overlying and underlying units in the stratigraphic column to possess stratigraphic significance; other chert-bearing zones in the limestone are characterized over large areas by uniformity in the amount, color, type of chert mass, size of nodules, and other properties. This indicates uniformity of silica supply over equally large areas, a condition that is more readily obtained from a uniform marine water source immediately adjacent to the planar deposition site than from some later movement of silica-bearing waters along or across beds of heterogeneous susceptibility to replacement. Conversely, chert may be absent or scarce in certain well-defined zones of wide stratigraphic distribution in an otherwise cherty sequence, such as the 10- to 20-foot zone near the base of the Joplin Member. This fact further suggests that the presence or absence of silica in the strata was determined at the time of sedimentation, for it is inconceivable that such a porous unit that has been readily leached in areas of ground-water movement in the mining field would not have been widely chertified along with the rest of the enclosing section if the introduction of silica had been effected by circulating ground waters after deposition of the complete Boone section.

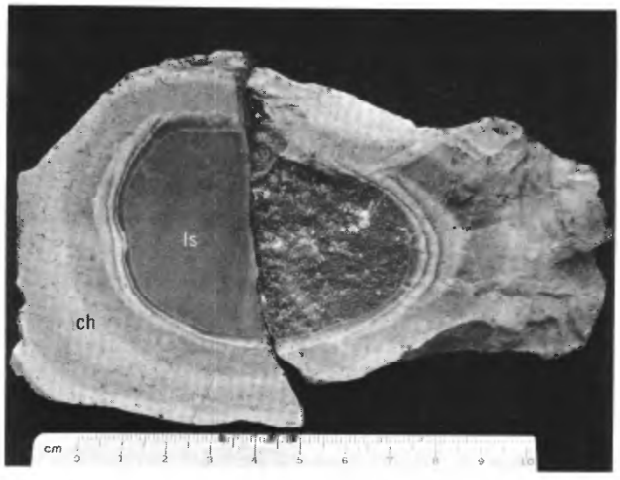

FIGURE 18.—Concentric banding in chert (ch) surrounding a nodular-shaped core of limestone (ls), from K bed, Mary Jane mine. Left side of specimen is sawed; right side shows natural break.

The amount of chert is roughly related to the stratigraphic horizon at which it occurs, but there is lateral and usually gradual variation in original content of a given bed from place to place. This is entirely aside from secondary variations brought about by solution of the limestone, though the two phenomena have not always been distinguished from each other. Local fluctuations in original chert content within one bed are independent of those in another; hence, a large proportion of chert is maintained everywhere in the Boone, although Giles (1935, p. 1832–1834) has indicated a gradual regional decrease in chert content from west to east in a 100-mile-long belt from northeast Oklahoma across part of northern Arkansas. Even though the content and character may vary locally in a given stratigraphic interval, there is likely to be an average condition that is widespread and characteristic of the interval, recurring beyond local small areas of aberration; hence the suitability of the chert for stratigraphic correlations.

The obvious relationship of the chert to the Boone stratigraphy and its widespread occurrence in the Mississippian limestones not only of North America but also of Europe and Asia were major difficulties that Giles (1935, p. 1860, 1866) found hard to reconcile with the theory of hydrothermal origin in post-Boone time.

We believe that the silica fixed in the chert and cotton rock was derived from the Mississippian sea nearly at the time that the enclosing limestone was precipitated. The replacement may well have occurred in the top few inches to 4 feet of sediment on the floor of the sea, in an early stage of diagenesis. Dapples (1959, p. 49), in discussing the deposition of chert in Niagaran strata of Illinois and Indiana, says,

> The summation of all the relationships involved point rather strongly to the development of chert not as an original precipitate during deposition, but as a product of diagenesis formed primarily during the stage of early burial. Precipitation of such chert must, therefore, have occurred from interstitial waters oversaturated with silica under the conditions which prevailed.

That the Early to Middle Mississippian was a period of high silica concentration in the midcontinental sea is a corollary of Goldstein and Henricks' (1953) conclusion that the silica in several specified Paleozoic formations of the Ouachita Mountain area, which include cor-

relatives of the Boone Formation (Branson, 1959, p. 119), was derived from submarine weathering of volcanic ash. An excess of silica is released when volcanic ash weathers to its normal alteration product, bentonite (Ross and Hendricks, 1945, p. 66–67); and such silica could be readily taken into solution by sea water to a maximum extent far in excess of the amount occurring in present-day sea waters (Krauskopf, 1956, p. 21–22). The form in which the silica was deposited would differ with the depositional environment. In a marine-limestone environment beyond the area of direct ash fall, the dissolved silica in the sea water that saturates the porous limestone on the floor of the sea could be an adequate source for the chert. The fact that the silica has been leached from material in contact with the sea water would assure that concentrations would be below the saturation point of amorphous silica, whereas replacement of a suspectible host rock is feasible from such a solution because of its oversaturation in silica with respect to quartz (Dapples, 1959, p. 49).

Although sponge spicules are commonly preserved in many of the Boone chert masses, their abundance varies widely, and they are not noticeable in other equally abundant cherts. There is no indication that the silica was initially precipitated in place, either by organisms or by some such inorganic process as a direct volcanic ash fall, to be later aggregated into nodules by dissolution and diffusion through the lime. As the limestone interstitial to the chert is virtually free of silica (Giles 1935, p. 1849–1853), such a theory would present difficulties in explaining complete resolution, whereas a single direct precipitation by replacement from unsaturated sea water would offer no such complication. Although there would be a tendency for a high concentration of silica at certain times to replace the lime floor over a wide area, the effectiveness might vary from place to place, accounting for the observed variation in the chert content at a given horizon.

Bastin (1933, p. 377, 381) and Fowler and others (1935, fig. 15a–3) have cited the occurrence of stylolites on chert-limestone contacts as evidence that the chert has replaced limestone later than the formation of the stylolite, and thus later than diagenesis and compaction of the limestone. As pointed out by Heald (1955, p. 108), this argument does not allow for the possibility of later stylolitic solution in the chert. Stylolites have been described in quartzite and sandstone where the two interpenetrating bodies are of roughly comparable composition (Tarr, 1916; Young, 1945; Heald, 1955), and stylolitic solution also occurs in calcareous sandstones between quartz and calcite grains (Heald, 1955, p. 108). Stylolites on the contact between Boone limestone and chert have a smaller wave length and amplitude than those commonly prevailing where only limestone is involved. This may well mean that the chert influenced the character of the stylolite and was thus earlier. Such stylolites occur on the borders of chert nodules or lenses where the amount of interstitial lime is considerable; and as the chert is much finer crystalline than sandstone and hence more soluble, it is not surprising that stylolites can develop on the contact between the limestone and chert phases. Many such stylolites were formed early and were broken and rotated during the tectonic brecciation.

In heavily leached ground where the limestone has been removed and only chert remains, a coarser type of stylolite develops between the chert residuals, in places along the broken edges of chert fragments. Such stylolites have a nearly horizontal attitude even where bedding has been destroyed by brecciation and rotation of the chert masses during preceding stages in the history of the rock mass; hence, the pressure of gravity would appear to be the major control in determing the orientation of the stylolite. These stylolites obviously are later than the chert, and also later than the tectonic brecciation.

Secondary Chert

Several authors (Fowler and others, 1935, p. 136–151; Laney, 1917; Buckley and Buehler, 1906, p. 30–31) have described two ages of chert in the Tri-State region, yet the criteria for distinction between the two ages have not been clear. Gregory and Agar (in Fowler and others, 1935) record the occurrence of older chert fragments in a later chert matrix and describe certain petrographic differences which they apply over large areas, leading to the conclusion that the later chert is more widespread in the Tri-State region than the earlier. Yet the differences between the two types as described are mostly of degree rather than kind, their mutual boundaries are commonly gradational, and there is no convincing evidence that the "later" type, as extrapolated beyond the breccia occurrences, is anything more than the less complete replacement stage of a single period of chertification.

Angular older chert fragments included in a later chert matrix of lighter color has been observed locally by us, but such material is insignificant bulkwise. Thus, although two ages of chert are obviously present locally, the differences are not great enough to allow recognition in unbroken sections. Color and luster variations within a single unbroken chert nodule or bed of primary origin commonly encompass a range greater than that recognized between the two chert generations in the brecci-

ated material, and petrographic distinctions in thin sections are equally uncertain.

Hence, other criteria have to be used in separating the two generations. The most significant is the macrotextures of certain chert and cotton-rock masses, present only in ground that has been intensely leached and condensed by underground solutions. In addition to the scattered breccias that demonstrate two ages of chert, such ground commonly contains chert in which the constituent units are a close-fitting mosaic of irregular-shaped interlocking masses in varying sizes that form a "headcheese" texture (fig. 19). Size of the chert units forming the mosaic generally ranges from a very small fraction of an inch to 1 inch across, but may be as much as 3 inches across. Thin and commonly indefinite dark selvage surfaces separating the units of the mosaic represent the clay fraction of the original limestone that was pushed laterally during the replacement and segregated at the edges of the separate chert nuclei. That this headcheese texture is a replacement feature and not a chert breccia modified by solution and condensation is proven by the local occurrence of undeformed crinoid stem casts in the mosaic and particularly in jasperoidized interstitial limestone in narrow gradational zones between solid headcheese masses and adjacent fossiliferous jasperoid. However, the headcheese masses may also enclose both broken and unbroken chert nodules of the older generation as well as less easily classifiable chert masses of indefinite shape and uncertain relative age. Contacts between the original chert nodules and the headcheese are commonly stylolites of coarser texture than those found between chert and limestone in less thoroughly leached ground.

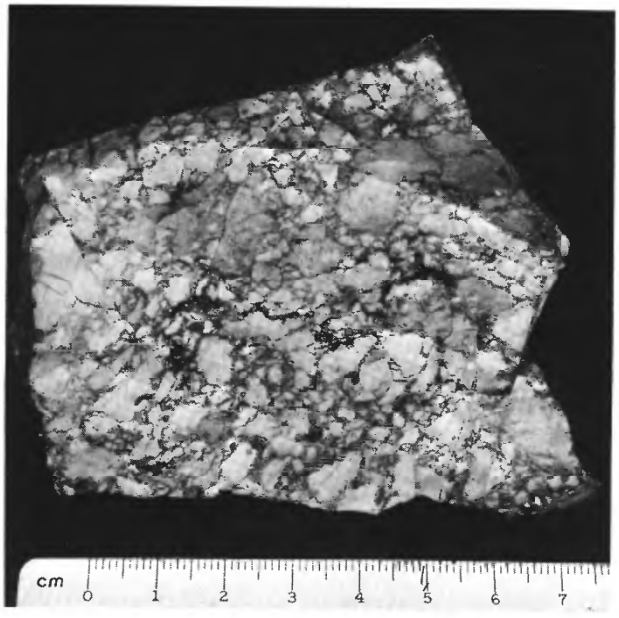

FIGURE 19.—Specimen from East Netta mine, showing "headcheese" texture that characterizes secondary chert formed by replacement of limestone in Joplin Member of Boone Formation during period of extensive leaching by ground waters. Some units, including the darker ones, are fragments of the earlier generation of chert.

In the headcheese mosaics, cotton rock forms the later generation as commonly as chert. Although the headcheese replacement chert and cotton rock is very similar to certain primary textures in basal L bed (p. 42), and some of it might have formed during the primary chertification, its common occurrence in leached ground in the mines and its absence in unaltered ground are presumptive evidence of its secondary origin. Although it usually occurs in indefinite bodies elongated approximately with the bedding, it may occur in cross cutting masses. The common association with underground pipes and pockets of the Pennsylvanian shale indicates that the known period of extensive underground solution following uplift at the end of Chester time and ending with the invasion of the Pennsylvanian sea was also the period when a certain amount of silica was redistributed in solution. The abundant chert and cotton rock of the earlier generation constituted an ample source. The amount of silica reprecipitated, however, was minor relative to that in the first generation. It is possible, also, that the Grand Falls chert, which is predominantly cotton rock over much of the Wyandotte quadrangle, was locally converted by these ground waters to the hard chert characteristic of much of the mining field. This redistribution of silica was concomitant with a far more extensive leaching of the limestone by the same waters, concentrating the less soluble primary chert and producing, by condensation, a pseudochertification that is more apparent than real.

Another mode of chert occurrence that is definitely secondary but of even less importance bulkwise is in crosscutting replacement veins, as observed in the tripoli bed at a quarry a little more than a mile northwest of Seneca. The veins are only a few inches thick at most and are localized on slickensides of negligible displacement. They are not persistent in either strike or dip.

All chert that is definitely secondary or suspected of being secondary is alike in being pale gray or pale brown to nearly white; most of it is opaque and has a rather dull luster. It is thus indistinguishable in petrologic detail from some of the primary chert.

It would at first appear that a study of churn-drill

cuttings could readily distinguish secondary chertification of the Boone from an apparent chertification wherein the increase in percentage of chert is due solely to dissolution of most or all of the associated limestone. Under the latter conditions, a decrease in the thickness of the strategraphic units might be expected. Such a decrease can be demonstrated in many of the drill holes, but in other examples involving all the Boone above K bed, no appreciable decrease has taken place. However, in this latter group, the coarseness of the chert cuttings in many of the drill holes and the leached appearance of some surfaces on the cuttings in even more drill holes is adequate evidence for the existence of voids in the section, showing that the ground was leached but failed to collapse. Evidently the closeset distribution of the chert beds in the Moccasin Bend Member commonly allowed enough points of contact in the skeletal network of chert to prevent any noticeable collapse when the limestone was dissolved out. On the other hand, the amount and distribution of chert in the Joplin Member are such that this unit presents no competence on loss of the limestone; thus, the unit almost everywhere is thinned if the chert has been concentrated by leaching.

ORE DEPOSITS

INTRODUCTORY SUMMARY

The position and importance of the Picher field within the broader Tri-State mining region have been indicated in the introduction to this report (p. 5–6). The Tri-State region is roughly 125 miles long east to west and 50 miles wide, and it contains several scattered centers of zinc-lead production. The Picher field lies at the west end of this mineralized region. Fringe deposits of the field were discovered southeast of Quapaw about 1901, but the main part of the field, lying 2–8 miles northwest of this town, was not found until 11 years later. Rapid development during World War I quickly brought the field to the forefront in zinc-lead mining, and peak production was reached in 1925. Moderately high production was sustained for most years until the late 1950's when the yield declined sharply. The total production through 1964, in terms of recoverable metal, has been about 7,283,000 tons of zinc and 1,766,000 tons of lead, valued at about $1.4 billion (table 4). The average grade of the ore has been moderately low, ranging mostly between 2 and 5 percent combined zinc and lead.

TABLE 4.—*Zinc and lead produced in the Picher field (Ottawa County, Okla., and Cherokee County, Kans.), 1904–64*

Subdistricts included are Lincolnville, Quapaw, Sunnyside, Picher-Cardin, Century, Commerce, Miami, and Melrose in Oklahoma; Baxter Springs-Blue Mound, Treece, and Melrose in Kansas. Figures for 1904–06 from A. J. Martin (1946, p. 29, 63); those for 1907–31 based on tables published in annual volumes of Mineral Resources of the United States; those for 1932–64 based on unpublished statistical charts furnished by U.S. Bur. Mines]

Year	Lead concentrates (galena)		Zinc concentrates (sphalerite)		Recoverable metal content [1]			
					Lead		Zinc	
	Short tons	Value [2]	Short tons	Value [2]	Short tons	Value [2]	Short tons	Value [2]
1904	150	$8, 250	633	$21, 245	112	$9, 789	317	$32, 841
1905	566	34, 450	2, 670	103, 480	422	40, 174	1, 354	159, 772
1906	669	51, 299	3, 242	124, 528	498	56, 374	1, 624	201, 051
1907	647	43, 644	[3] 3, 159	120, 071	500	53, 000	1, 495	176, 410
1908	2, 234	118, 253	[3] 10, 033	249, 674	1, 726	144, 984	4, 404	413, 976
1909	4, 300	223, 131	[3] 16, 622	569, 200	3, 319	285, 434	7, 665	827, 820
1910	3, 634	187, 861	13, 976	447, 043	2, 798	246, 224	6, 305	680, 940
1911	3, 177	170, 729	[3] 10, 642	330, 186	2, 416	212, 608	4, 963	565, 782
1912	4, 257	231, 678	[3] 11, 881	484, 429	3, 286	295, 740	5, 627	776, 526
1913	7, 807	402, 927	24, 097	766, 200	6, 039	531, 432	11, 649	1, 303, 568
1914	9, 402	443, 543	[3] 28, 367	926, 778	7, 329	571, 662	13, 990	1, 426, 980
1915	9, 058	494, 524	28, 280	1, 901, 490	6, 934	651, 796	14, 191	3, 519, 368
1916	15, 206	1, 275, 761	54, 932	4, 109, 565	11, 777	1, 625, 226	28, 498	7, 637, 464
1917	33, 770	3, 401, 926	171, 726	11, 611, 675	26, 624	4, 685, 824	92, 339	18, 837, 156
1918	77, 487	6, 889, 080	341, 175	17, 321, 065	60, 924	9, 016, 752	183, 434	33, 384, 988
1919	81, 290	5, 524, 106	413, 418	17, 892, 434	63, 427	7, 357, 532	219, 792	32, 089, 632
1920	101, 285	9, 560, 901	502, 134	22, 610, 299	79, 755	12, 760, 800	270, 610	43, 838, 820
1921	74, 580	3, 949, 045	278, 331	6, 344, 770	59, 977	5, 397, 930	149, 623	14, 962, 300
1922	108, 510	8, 240, 542	482, 970	16, 528, 301	85, 628	9, 419, 080	260, 119	29, 653, 566
1923	107, 496	10, 255, 061	633, 035	25, 656, 673	84, 045	11, 766, 300	332, 224	45, 182, 464
1924	113, 363	12, 142, 523	690, 809	28, 502, 120	88, 074	14, 091, 840	361, 073	46, 939, 490
1925	130, 410	15, 324, 698	749, 254	38, 303, 908	100, 838	17, 545, 812	387, 002	58, 824, 304
1926	124, 361	13, 226, 619	744, 028	34, 567, 144	95, 832	15, 333, 120	382, 683	57, 402, 450
1927	99, 524	8, 689, 985	591, 447	22, 945, 385	76, 404	9, 626, 904	303, 298	38, 822, 144
1928	87, 238	7, 054, 366	527, 495	19, 355, 536	67, 406	7, 819, 096	271, 116	33, 076, 152
1929	91, 087	7, 687, 831	562, 371	22, 091, 618	69, 699	8, 782, 074	290, 375	38, 329, 500
1930	45, 492	2, 994, 261	394, 459	12, 044, 167	34, 291	3, 429, 100	204, 363	19, 618, 848

See footnotes at end of table.

TABLE 4.—*Zinc and lead produced in the Picher field (Ottawa County, Okla., and Cherokee County, Kans.), 1904–64*—Continued

Year	Lead concentrates (galena)		Zinc concentrates (sphalerite)		Recoverable metal content [1]			
					Lead		Zinc	
	Short tons	Value [2]	Short tons	Value [2]	Short tons	Value [2]	Short tons	Value [2]
1931	24, 565	1, 115, 883	218, 689	5, 119, 691	18, 990	1, 405, 260	115, 569	8, 783, 244
1932	21, 130	765, 715	167, 725	3, 066, 363	16, 461	987, 660	89, 686	5, 381, 160
1933	30, 820	1, 378, 300	248, 933	6, 507, 023	23, 643	1, 749, 582	131, 761	11, 067, 924
1934	30, 222	1, 183, 580	276, 887	7, 527, 784	23, 250	1, 720, 500	145, 900	12, 547, 400
1935	44, 715	1, 894, 278	344, 927	9, 916, 502	34, 035	2, 722, 800	182, 300	16, 042, 400
1936	48, 545	2, 498, 168	391, 383	12, 291, 640	36, 778	3, 383, 576	206, 974	20, 697, 400
1937	59, 906	4, 181, 830	403, 783	16, 752, 621	45, 799	5, 404, 282	214, 080	27, 830, 400
1938	47, 461	2, 467, 059	335, 927	10, 340, 762	36, 200	3, 330, 400	182, 463	17, 533, 728
1939	53, 654	3, 163, 728	379, 435	13, 075, 004	41, 396	3, 891, 224	206, 598	21, 486, 192
1940	43, 290	2, 561, 072	400, 647	16, 874, 966	33, 131	3, 313, 100	217, 028	27, 345, 528
1941	51, 301	3, 452, 758	429, 660	21, 569, 025	39, 391	4, 490, 574	233, 173	34, 975, 950
1942	41, 942	4, 141, 239	369, 043	25, 585, 281	32, 146	4, 307, 564	198, 931	37, 001, 166
1943	38, 536	4, 652, 150	310, 980	30, 836, 571	28, 851	4, 327, 650	166, 850	36, 039, 600
1944	29, 952	3, 532, 942	275, 220	27, 906, 030	22, 817	3, 650, 720	148, 125	33, 772, 500
1945	25, 813	3, 209, 989	206, 580	22, 402, 691	19, 043	3, 275, 396	111, 486	25, 641, 780
1946	25, 880	4, 213, 144	206, 764	23, 790, 583	19, 800	4, 316, 400	111, 558	27, 220, 152
1947	27, 805	5, 299, 841	164, 574	17, 586, 767	21, 104	6, 077, 952	88, 634	21, 449, 428
1948	32, 392	7, 482, 361	143, 429	12, 539, 196	24, 294	8, 697, 252	76, 409	20, 324, 794
1949	38, 785	7, 249, 558	134, 513	10, 382, 456	28, 767	9, 090, 372	71, 895	17, 829, 960
1950	39, 073	5, 993, 270	137, 275	12, 792, 548	29, 910	8, 075, 700	73, 701	20, 931, 084
1951	34, 468	7, 282, 897	152, 853	18, 944, 275	25, 474	8, 814, 004	82, 333	29, 969, 212
1952	28, 010	5, 643, 233	148, 488	17, 363, 541	20, 887	6, 725, 614	80, 229	26, 636, 028
1953	16, 608	2, 579, 916	90, 541	6, 594, 882	12, 649	3, 314, 038	48, 917	11, 250, 910
1954	24, 394	4, 110, 406	123, 340	8, 104, 829	18, 237	4, 996, 938	62, 281	13, 452, 696
1955	26, 917	4, 721, 589	129, 978	9, 977, 920	19, 624	5, 847, 952	69, 154	17, 011, 884
1956	28, 101	5, 180, 293	106, 135	9, 173, 252	19, 985	6, 275, 290	56, 180	15, 393, 320
1957	15, 901	2, 922, 323	56, 891	4, 599, 858	11, 440	3, 271, 840	30, 810	7, 147, 920
1958	7, 041	931, 441	18, 001	1, 093, 365	4, 991	1, 167, 894	9, 688	1, 976, 352
1959	1, 607	211, 244	4, 061	282, 285	1, 082	248, 860	2, 066	475, 180
1960	3, 098	283, 818	8, 877	657, 838	1, 717	401, 778	4, 449	1, 147, 842
1961	3, 243	352, 243	10, 666	716, 541	2, 429	500, 374	5, 594	1, 286, 620
1962	4, 890	480, 412	25, 564	1, 771, 466	3, 680	677, 120	13, 956	3, 209, 880
1963	5, 719	604, 378	30, 762	2, 270, 865	4, 219	911, 304	16, 753	3, 853, 190
1964	5, 333	733, 391	31, 228	2, 731, 701	3, 966	1, 039, 092	16, 824	4, 576, 128
Total	2, 298, 117	225, 101, 443	13, 774, 945	697, 075, 106	1, 766, 266	270, 166, 669	7, 282, 668	1, 109, 973, 264

[1] Allowance has been made for smelting losses of both lead and zinc.
[2] In comparing the values of metal and concentrates it should be borne in mind that the value given for the metal is calculated from the average price for all grades, whereas the value given for the concentrates is that actually received by the producers.
[3] Includes a small quantity of silicate and carbonate.

The ore deposits are bodies of ore and gangue minerals that replace limestone in favorable stratigraphic zones chiefly in the Boone Formation. Most of the ore has come from the Joplin Member and the middle of the Baxter Springs Member (respectively, M and K beds of Fowler and Lyden, 1932), some has come from the underlying Grand Falls Chert Member, some from G, H, and E beds of the overlying Moccasin Bend Member, and a little has come from other parts of the Boone and from overlying strata of the Chester Series. Where they are not mineralized or leached, most of these ore-bearing beds are limestone that contains from 20 to 60 percent chert in beds and nodules. The Grand Falls Chert Member, however, is composed dominantly of chert layers, with thin irregular sheets of limestone between the chert layers.

Before deposition of the overlying Krebs Group shale of Pennsylvanian age, considerable limestone was leached from the Boone and Chester beds. Caves developed in places, but much of this solution caused simultaneous collapse and thinning of these beds. This collapsed ground contains brecciated chert and a moderate amount of clay and shale, some of which is residual from the limestone, and some is Krebs shale that fills caves and was introduced through channelways open to the surface on which the Krebs shale was being deposited. This leached ground is the locus for later ore mineralization, which largely replaced the remaining limestone. However, not necessarily all ore-bearing ground was leached in that period.

The strata in the Picher field are on the whole nearly flat lying, or are gently flexed in an irregular pattern. Superimposed on this broad simple structure, however, are the Miami trough, crossing the west side of the field, and the Bendelari monocline, which is nearly at right angles to the trough. These are linear tectonic features

that sharply bend or break the strata in and adjacent to them. Both reflect movement along deep-seated fractures in the Precambrian crystalline rocks. Other pronounced structural features include many sharply defined slump basins and slump pipes that probably resulted from collapse of beds into leached ground at different stratigraphic levels in the Boone and underlying strata. Probably concealed structural breaks of small amplitude localized the leaching. The gentle flexures of medium size and irregular pattern are of uncertain or mixed origin, though probably many were of tectonic inception.

The Miami trough influenced the localization and shape of some ore bodies along it, but not of all (pl. 1). There is, in addition, a large-scale alinement of the ore deposits of the field along northwest and northeast lines. Although vague in parts of the field, this alinement is pronounced in other parts (pl. 1). The northwest trend is roughly parallel to the better defined sector of the Bendelari monocline northwest of the Miami trough. The northeast trend is in part nearly parallel to the Miami trough, but in part at an appreciable angle to it. The alinements may be controlled by deep crustal fractures in the same general systems with the trough and monocline fractures, but not always closely parallel to them. The slump basins and slump pipes preceded the ore but have little relation to the ore deposits, except indirectly in that they are more abundant near the Miami trough. Some of the mineralized fracture zones that have been stoped are crudely parallel to structure contours defining the gentler flexures (pls. 5–10).

Structural movement on the Miami trough and Bendelari monocline took place in the interval between the deposition of the Chester and that of the Krebs, and was accentuated after deposition of the Krebs. Brecciation of the chert in the Boone was concomitant with plastic deformation of the enclosing limestone during slight crustal warping, and probably was extensive in the first-mentioned interval of structural deformation. This brecciation may well have been responsible for opening up the ground to solution by ground waters in the erosion interval preceding deposition of the Krebs strata. The post-Krebs structural movement may have contributed further brecciation. The latest deformation, however, apparently came after the rock had acquired considerable rigidity, so that joint systems were developed across the previously deformed Boone strata. These joint systems are vertical or steeply dipping, curvilinear, discontinuous, wide in places, and narrow in others. The mineralization followed, and was in part controlled by the joint systems, although there was also much spreading of ore through permeable ground prepared during the interval of pre-Krebs leaching. Some of the ore deposits replaced the remaining limestone in the ground broken by the joint systems where these crossed favorable ore beds. Elsewhere, conditions were unfavorable for ore deposition along the joint systems but became favorable in peripheral zones in the ore beds outward from the joint systems. Further leaching and condensation of the ore beds accompanied the mineralization, and additional jointing developed in the collapsed strata above the ore deposits, particularly those of linear trend.

Nearly all ore bodies are tabular masses whose horizontal dimensions exceed the vertical. Some are blanket-like bodies, dominantly irregular or lobate in plan but tending to be slightly elongate and curved; these grade into others, called runs, which are narrow, conspicuously elongate, and usually curvilinear. Many of the runs tend to form closed but irregular-shaped circles around barren cores. The sides of most ore bodies are assay boundaries; but in many cases the decrease in grade is rather abrupt laterally, and the limits are rather sharp. Most ore bodies are largely confined to a definite stratigraphic interval, so the tops and bottoms of these are therefore crudely parallel. Stopes in bodies of this type are commoniy 10–20 feet high. In other places two or more stratigraphic units contain ore bodies that are superposed or partly overlap. Where they are stratigraphically adjacent, they are mined together, and in such places the stopes may be 50–100 feet high. If the ore-bearing units are separated by much waste rock, they are mined by separate stopes. Mineralization in the central part of the Picher field seems to have been a little more intense than on the margins; hence, the occurrence of superposed ore bodies in two or more stratigraphic units is somewhat more common in the central part of the field.

Structural deformation had generally shattered the chert contained in the ore beds prior to mineralization, so that much of the ore is in the matrix of a chert breccia. The limestone that originally formed this matrix was either removed by leaching or was entirely replaced by the ore and gangue minerals.

The ore consists of sphalerite, galena, dolomite, and jasperoid, with an unreplaced residuum of chert. Accessory metallic minerals are chalcopyrite, enargite, luzonite, marcasite, and pyrite; all except marcasite are sparse. Considerable calcite and locally a little quartz and barite occur in the ore, and large calcite crystals are present in caves adjacent to ore bodies. The zinc to lead ratio for the ore, based on the total production of the field, is about 4.1:1. Individual mine outputs may vary, however, from wholly zinc to predominantly lead

356–188 O—69——7

in a few mines of minor output. Owing to the intimate admixture of ore minerals with the gangue and country rock, all ore has to be milled.

Paragenetic relations are complex in detail, owing to partial overlap and repeated periods of crystallization of some minerals, but the major part of each mineral was precipitated in the following sequence: Dolomite, jasperoid, sphalerite, chalcopyrite, galena, marcasite and pyrite, enargite and luzonite, and calcite. The dolomite replaced the limestone to give a coarse-grained massive gray rock ("gray spar"). It also occurs abundantly as a pink filling of veinlets and irregular pockets or as a lining of cavities in the replaced limestone and adjacent chert masses, commonly with well-formed crystals on the accompanying drusy surfaces ("pink spar"). Jasperoid is almost entirely a massive replacement of limestone or of the gray spar dolomite. Sphalerite and galena replace the limestone and earlier gangue minerals as disseminated crystals which may become so abundant as to constitute a massive crystalline replacement. They also occur like, and may accompany, the pink spar dolomite in bedding seams, crosscutting veinlets, vugs, cave linings, and as a coating or interstitial filling of the residual chert masses left by leaching of the limestone; in all these occurrences, crystals of varying but commonly large size are abundant in the associated cavities. The chalcopyrite, enargite, and luzonite occur solely or predominantly in the cavities. Marcasite and pyrite are most conspicous in this same occurrence, but also to some extent replace the jasperoid and gray spar dolomite, partly as disseminated microscopic crystals. Calcite fills late cavities, but it also occurs in veinlets and irregular replacement masses in the mineralized ground.

Many ore bodies show horizontal zoning that involves both the gangue and ore minerals. Dolomite was the first mineral introduced in the ore-forming event, and it replaced the limestone of the pertinent ore bed over areas that are commonly several hundred feet across. Jasperoid mainly replaced the limestone surrounding the dolomite masses, but it also partly replaced some of the dolomite, especially around the fringes. Beyond the ring of jasperoid the limestone in some places was largely to completely leached, leaving a residuum of chert nodules, a little clay, and limestone in varying proportions. Beyond this conspicuously leached ground, commonly called boulder ground or rubble ground, the limestone is unleached or only partly leached, and the stratigraphic unit is generally much thicker than where mineralized. The sphalerite and galena were precipitated chiefly along the outer edge of the dolomite core; this edge is, therefore, the locus of the richest ore runs. Sphalerite tends to concentrate on the dolomite side of the run; and although the concentration decreases abruptly toward the dolomite core, enough sphalerite may persist in some places to constitute low-grade ore through much of the core. Galena occurs with sphalerite in the ore run, but it is relatively more abundant on the outer or jasperoid side of the run. The copper-bearing sulfides are associated exclusively with the dolomite, whereas marcasite and calcite are relatively more abundant with the jasperoid or in the boulder ground, though they also occur in the dolomite. In the boulder ground the chert nodules are commonly coated with a film of jasperoid and scattered small crystals of sphalerite or galena, but usually the grade is too low for mining.

The ores are believed to have formed from hydrothermal solutions derived from a deep-seated magma. These solutions probably rose along the Miami trough and other deep-seated fractures in the Precambrian basement rocks, and then along these fractures and other channelways through the lower Paleozoic strata until they came to the favorable beds of limestone and chert of Mississippian age, where in part they spread laterally to form the tabular and zoned ore bodies. The sites of ore deposition had already been partly prepared by mild deformation—fracturing, chert brecciation, and some collapse of the beds due to partial leaching of the limestone—before accumulation of Pennsylvanian shales. Mineralization occurred at some later but unknown period, possibly during the Cretaceous or perhaps even a little later.

The deposits are mined through vertical shafts. Most deposits lie from 100–300 feet below the prairie level; however, some in the eastern part of the field are shallower, and some in the Miami trough may lie at depths as great as 480 feet. Except in the eastern part of the mining field, the shafts are collared in the Pennsylvanian black shale that blankets the ore-bearing Chester and Boone strata to varying depths, generally less than 100 feet.

Many rich deposits were found early in the history of the field. For example, a little more than 400 tons of ore treated in 3 successive days' run of the Golden Rod No. 2 mill in December 1917, yielded 37 percent of concentrates in a ratio of about 2 sphalerite to 1 of galena. These concentrates ran 62 percent zinc and 82 percent lead, respectively, which is somewhat above standard market grade. Expressed in terms of metal, the mill recovered about 15 percent zinc and 10 percent lead from the crude ore. As might be expected, larger samples averaged lower in grade, though 1 year's run of 49,000

tons in the early history of the Malsbury mine yielded more than 24 percent of concentrates, equivalent to 12 percent zinc and 3 percent lead.

Such deposits have now been largely mined out, and the mining industry has turned to the lower grade blocks of ground that were passed by in the early days. The grade of ore in recent years has been influenced mainly by the economics of mining. Because of the shallow depth of the deposits, their amenability to large tonnage production, their mineralogic simplicity with resultant ease of milling, and especially the technical efficiency of the Tri-State mining industry, the costs per ton of concentrates produced are among the lowest in the country. Hence, the grade of ore mined has been lower than in any other predominantly zinc district except the adjacent Missouri part of the Tri-State region. Table 5 shows that in 1946 when price-cost relations were most favorable to Tri-State mining, the crude ore mined in the Picher field averaged only 1.99 percent of combined zinc and lead metal recovered in the concentrates.

TABLE 5.—*Approximate tenor of zinc-lead ore mined in Picher field* [1]

[Compiled by Gwendolyn W. Luttrell and E. T. McKnight from unpublished statistical tables furnished by the U.S. Bur. Mines; does not include tailings]

Year	Tons crude ore	Percent combined metals [1]	Percent lead [1]	Percent zinc [1]
1904–10	Data incomplete			
1911	224, 450	3. 56	1. 11	2. 45
1912	290, 200	3. 31	1. 16	2. 15
1913	579, 200	3. 30	1. 07	2. 23
1914	760, 670	3. 02	. 99	2. 03
1915	877, 700	2. 59	. 81	1. 78
1916	1, 304, 300	3. 49	. 92	2. 57
1917 [2]	3, 418, 700	3. 72	. 77	2. 95
1918	6, 464, 400	4. 08	. 97	3. 11
1919	7, 201, 270	4. 36	. 90	3. 46
1920	8, 164, 600	4. 88	1. 00	3. 88
1921	3, 827, 900	6. 22	1. 58	4. 64
1922	8, 032, 000	4. 60	1. 09	3. 51
1923	9, 943, 450	4. 34	. 86	3. 48
1924	11, 324, 200	4. 29	. 79	3. 50
1925	13, 623, 400	3. 95	. 78	3. 17
1926	13, 934, 100	3. 75	. 69	3. 06
1927	10, 565, 000	3. 71	. 73	2. 98
1928	8, 482, 400	3. 92	. 73	3. 19
1929	10, 061, 500	3. 70	. 62	3. 08
1930	6, 200, 600	3. 94	. 70	3. 24
1931	3, 125, 300	4. 71	. 69	4. 02
1932	1, 754, 200	4. 51	. 62	3. 89
1933	2, 896, 100	4. 44	. 69	3. 75
1934	3, 701, 600	4. 34	. 73	3. 61
1935	4, 086, 100	4. 48	. 66	3. 82
1936	4, 767, 900	4. 26	. 70	3. 56
1937	5, 859, 300	3. 89	. 76	3. 13
1938	5, 065, 400	3. 91	. 74	3. 17
1939	5, 141, 800	4. 37	. 68	3. 69
1940	5, 919, 350	3. 82	. 56	3. 26
1941	7, 117, 320	3. 61	. 56	3. 05
1942	7, 017, 270	3. 14	. 46	2. 68
1943	7, 363, 010	2. 51	. 39	2. 12
1944	6, 745, 620	2. 36	. 34	2. 02
1945	5, 641, 120	2. 14	. 34	1. 80
1946	6, 295, 020	1. 99	. 31	1. 68
1947	4, 812, 790	2. 23	. 44	1. 79
1948	3, 818, 690	2. 68	. 64	2. 04
1949	4, 020, 090	2. 59	. 72	1. 87
1950	4, 377, 700	2. 50	. 69	1. 81
1951	5, 272, 540	2. 13	. 47	1. 66
1952	5, 153, 710	2. 09	. 41	1. 68
1953	2, 971, 473	2. 26	. 43	1. 83
1954	3, 872, 570	2. 26	. 48	1. 78
1955	4, 107, 145	2. 33	. 49	1. 84
1956	3, 429, 723	2. 41	. 59	1. 82
1957	1, 833, 542	2. 51	. 64	1. 87
1958	611, 556	2. 59	. 83	1. 76
1959	31, 750	6. 02	2. 09	3. 93
1960	51, 972	7. 71	1. 20	6. 51
1961	180, 331	4. 62	1. 31	3. 31
1962	474, 219	4. 06	. 79	3. 27
1963	612, 862	3. 74	. 70	3. 04
1964	691, 342	3. 28	. 58	2. 70

[1] Figures represent metal content of crude ore only insofar as it is recovered in the concentrates at first milling; additional recovery from tailings not included. The figures are, therefore, somewhat lower than the true grade of the crude ore.

[2] The figures for 1917 are for only the Oklahoma part of the field, which had accounted for all the production in prior years. Although there was initial production from the Kansas part of the field in 1917, some of the record has been lost. The production of crude ore was apparently not more than 6 or 7 percent of that from Oklahoma, and this probably would not change the grade of ore for the whole field materially from that shown for the Oklahoma part.

LANDOWNERSHIP PATTERN

Before ore was discovered, the land was privately owned agricultural land in rectangular-shaped plots that were fractional parts of square mile sections. The average plot was the quarter section of 160 acres, though some holdings exceeded this. Mining tracts have reflected these preexisting land subdivisions. The mine unit is most commonly the quarter-mile square of 40 acres, though some tracts, particularly in the earliest worked fringe areas, were subdivided to 10 acres, or even less, and others, in multiples of 40 acres, may be more than 300 acres. As the ore bodies are nearly flat, there are no apex problems. The operators mine to the vertical planes that mark the extension of the land boundary lines in depth. Mining rights are leased from the landowners who are paid a percentage royalty of the gross mineral sales from the tract. In Oklahoma, landownership was originally vested in individuals of the Quapaw Indian tribe, some of whom have profited handsomely from the royalty payments. A common practice in the history of the field has been for the holder of the first lease, whether an individual, royalty company, or mining company, to sublease at an increased royalty rate to the actual operator or other middle man. Particularly in the early days of the field,

royalties were all too commonly pyramided by this method to an excessive degree, which interfered with development of the field except in periods of high mineral prices. Most leases have been for comparatively short term and have usually required active operation to avoid forfeiture.

History of Mining

Ore was first discovered in 1901 about 1½ miles northeast of Lincolnville, Okla., by churn drilling on Abrams land (SE¼ sec. 30, T. 29 N., R. 24 E.) and on Julia Whitebird land (SW¼SW¼ sec. 29). In mid-1902, an exploratory shaft was sunk to ore at a depth of about 50 feet on one of the Abrams drill holes. The vigorous exploration that was stimulated by this initial discovery soon found ore on adjacent tracts. Small mills were built, and the first recorded output of sulfide concentrates from the Lincolnville area was made in 1904. Churn drilling of the area showed that the main deposits were in "blanket ground" at a depth of only 50–150 feet. In the 3 years ending with 1906, the production from the several shafts amounted to 6,545 tons of zinc sulfide concentrates and 1,385 tons of lead sulfide concentrates, at a total value of nearly $350,000. The Lincolnville deposits were comparatively short lived, however, with peak production in 1909. A few deposits here cropped out at the surface, and zinc silicate that had formed in the soil had, before discovery of the sulfides, been plowed up in a field and cast aside in ignorance of its identity. Much of this ore was later shipped when its true nature was recognized.

Zinc sulfide was found in cuttings from the town well at Quapaw, 2 miles northwest of Lincolnville, in or just before 1907. The next discoveries were made in another group of fringe deposits at the south side of the field, at what is now Commerce (formerly Hattenville). Following a churn-drilling campaign started here in 1905 by Robinson & Coleman (precursor of Commerce Mining & Royalty Co.), ore was reached by the first shaft in 1907. In the following year the production from this and several other shafts that had been put down nearby was more than double that in the Lincolnville area. In the early days, the Commerce mines were handicapped by a high royalty rate, by a large volumn of water that made pumping costs excessive, by high concentration of hydrogen sulfide in the ground, and by excessive tar in the ores, which complicated the milling problems and led to penalties in marketing the concentrates. Yet in spite of these handicaps, the ore was so much richer in comparison to other Tri-State mines that development of the area progressed rapidly. The early mines were comparatively shallow, at depths of 90–130 feet; but as the ore was followed northwestward in succeeding years, the depth increased to as much as 250 feet by 1911, and eventually to 385 feet. As the field was extended, the water and gas handicaps were greatly reduced through lowering of their initial head, and the amount of tar in the ore deposits decreased.

The exploration stimulated by the richness of the ores at Commerce and by the recognition of a northeasterly trend in some of these ore runs (see pl. 1) finally led to the discovery of the main part of the field in 1912. Although no mines were opened in the heart of the field that year, the drilling was extensive as far as the Kansas State line, and at least one important discovery was made in Kansas, on the Scott Jarrett land a few thousand feet north of the State line just west of the Blue Mound. Beginning in the following year, shafts were sunk and mills were built at an accelerating rate in the new part of the field. The first new mines were in the south, adjacent to the Commerce area.

A very important drilling discovery was made in the summer of 1914. The Picher Lead Co. had been exploring a large acreage between Commerce and the Kansas line, with results so discouraging it was decided to abandon the project. In attempting to get back to Joplin, a drill rig belonging to Dick Blosser got mired in a shallow prairie slough between Tar and Lytle Creeks, less than a mile south of the Kansas line. As there was going to be some delay before the rig could be extricated from the mud, and as the spot was within the large tract under lease to the company, the driller sought and was granted permission to put down a hole in order to retrieve some of the expense that had befallen him as a result of the mishap. On August 2 he struck ore of such richness as to change the decision of the company about abandoning the field (Eng. Mining Jour., 1943). Additional holes proved a rich ore body on the Crawfish land, on which this hole was drilled, and on several adjacent allotments of Indian land. By late 1915 the producing area of the field had expanded to include the Whitebird mine on one of these adjacent tracts, and the Crawfish, Bingham, and Netta mines began production shortly thereafter.

At the end of 1917 the Oklahoma part of the field was fairly well defined by producing mines, the Kansas deposits were largely outlined by new shafts or drill holes, and in that year the initial output from the Kansas side of the State line had been made from the Barr, Blue Mound, and Blue Diamond mines. Table 4 shows that the production of zinc concentrates virtually doubled in 1916 over the preceding year, tripled in 1917, doubled in 1918, and increased at a large, though some-

what less spectacular, rate in the next 2 years; the increase was from 28,000 tons in 1915 to 502,000 tons in 1920.

This phenomenal expansion was made possible by the multitude of mining companies operating in the field, each acting independently of the others in a favorable market stimulated by World War I. The landowners could award leases freely to any company that they might choose. Because of the shallowness of the deposits, the relative ease with which they could be mined and milled, and the wide technical experience gained from earlier operations in adjacent subdistricts of the Tri-State region, many relatively small but efficient mining companies were organized that required only moderate capital for operation on a 40-acre tract; the number of such tracts available was large. Hence, the average mining company tended to be small to moderate in the size of its operations. However, several major well-known companies or their subsidiaries entered the field on a modest scale. The St. Louis Smelting & Refining Co. (National Lead Co.), U.S. Smelting, Refining & Mining Co., Miami Zinc Syndicate (Butte & Superior Mining Co.), and Chanute Spelter Co. (American Metals Co.) became active in 1916, the United Zinc Smelting Corp. in 1917, and Federal Mining & Smelting Co. in 1918. Some of these abandoned the field after a short period, others remained as relatively small operators, and still others, particularly the St. Louis Smelting & Refining Co. and the Federal Mining & Smelting Co., expanded their operations later to sizable proportions. It was estimated in 1918 that there were 230 mills built or under construction in the Oklahoma part of the field, many having come secondhand from the older Missouri fields; and there were at least 20 more in the newer part of the field in Kansas. These figures can be taken as a rough measure of the number of individual operations. Some companies carried on multiple operations in areas that were not contiguous. Each operation ran the complete gamut from development drilling to the milling and marketing of concentrates, controlled only by the economics of mining within its pertinent unit. With the profits so widely dispersed, each unit tended to produce at maximum rate and efficiency. But to avoid forfeiture of leases that commonly obligated the operators to a stated minimum rate of output, there was also some forced production in excess of economic production, particularly after the price of zinc fell in early 1917. A large turnover in operating companies from year to year was inevitable as leases changed hands or were subdivided or regrouped under different operations, and the names of mines were constantly changing in a most bewildering fashion. As ore bodies commonly extend from one tract to the next, the concept of what constitutes a "mine" in the field is rather artificial.

The landownership pattern and leasing system prevented any domination of the field by a single company or small group of companies. Nevertheless, there were two mining companies that, in the initial development of the field, stood out above the others in the size and diversity of their operations. A parent company of the Commerce Mining & Royalty Co., with capital raised largely in Miami, the county seat of Ottawa County, had originally discovered and opened up the south edge of the field at Commerce (Hattenville). As the field grew, this company acquired control of a large mineral acreage, particularly through the western part of the field in Oklahoma, but also at one time or other, in scattered tracts throughout the field in both States. Initial activity, under the earlier name of Miami Royalty Co., was largely devoted to subleasing of its leased lands to other operators at an increased royalty rate over that paid on the first lease to the land owners, but the company soon became interested in active mining. The other leading operator, Picher Lead Co., originally of Joplin, which later merged with Eagle White Lead Co. to become the Eagle Picher Lead Co., leased 3,000 acres in Oklahoma, mainly just south of the Kansas line but with smaller segments north and northeast of Commerce. Exploration was started in 1913. The town of Picher grew on this leased land as a result of the rich discoveries and mine development. This company also subleased many of its properties to other companies from time to time.

Several other companies also stand out in the early history of the field. An association of Joplin and Miami men, under the company name of Church & Mabon, had taken part in the early development at the north end of the Commerce area, starting about 1910. When the main part of the field was discovered, this association gained control of a large acreage in the southern part of the field about 1½ miles southeast of Cardin, and under the new name of Welch Mining Co., proceeded to develop the ground. By mid-1917 the output of this company rivaled that of the two leading companies. At that time the leases on 200 acres southeast of Cardin were sold to the Skelton Lead and Zinc Co., which maintained the group of properties intact for many years during which the Skelton company was usually among the leading producers of the field.

The Golden Rod Mining & Smelting Co. in 1917 purchased several adjacent developed properties totaling 320 acres in the heart of the field, east of Cardin, including the old McConnell mine (now part of Kenoyer mine), which was the first property developed

in the Cardin area, as it had a shaft to ore in 1913. The company name was carried over from the Golden Rod Mining Co., an entirely different association, which in the preceding year had developed one of the purchased properties. In succeeding years, the output from the several mills on this block of land was an important segment of the total production from the field. The company later merged with a lead-products manufacturing concern to become the Evans-Wallower Lead Co., and later, Evans Wallower Zinc, Inc. The block near Cardin remained virtually intact until late in the history of the field, though additional holdings of the original company and its successors in other parts of the Picher field were bought and sold independently of the Cardin block.

The Underwriters Land Co., one of the oldest and largest operators in the old Joplin area, and the Admiralty Zinc Co. became active in the Picher field in 1915, and the latter company made its initial production in that year. In the following year the O. M. Bilharz Mining Co. entered the field with the purchase of two developed properties (Bluebird and Shorthorn). During the period of rapid development and accelerating production from 1916 to 1920 inclusive, these three companies also ranked among the 10 leading producers. The Admiralty and Bilharz companies remained steady producers for many years, although their relative importance was not so great later as during the early years. The Underwriters Land Co. lost its identity in the mid-1920's through a merger that produced an even more prominent successor company.

Because of the flatness of the terrain, transportation lines were no problem in the development of the field. The Oklahoma, Kansas, & Missouri Interurban Railway, later called the Northeast Oklahoma Railroad Co., which in 1908 had been constructed over the 4 miles from the main line of the St. Louis & San Francisco at Miami to Hattenville, was extended in 1916 to the new mining towns of Cardin and Picher, and spurs were extended to various mines as fast as developments warranted them. Another railway, the Miami Mineral Belt, was laid in 1917 from the main line at Quapaw to Cardin and Picher, and was extended in the next few years to Baxter Springs and also to much of the mining area. A third line, the Southwest Missouri Electric Railway, was completed in 1918 from Galena, Kans., to Picher.

Despite slumps in metal price that began in 1916 and were accentuated in the latter half of 1917, development of the field was carried forward largely on its own momentum and coincident with generally falling prices until a temporary peak of production was reached in 1920. Because of the number and type of operations, the adverse economic conditions during this period merely slowed the rate of increase in production without reversing its direction. Both zinc and lead prices reached lows in 1921, and the overall production of the field was at last curtailed drastically in that year. Many good mines were shut down for months, or even more than a year.

The setback was only temporary. With recovery in price during the next few years, the alltime peak of production was reached in 1925 when 749,000 tons of zinc concentrates and 130,000 tons of lead concentrates, equivalent to 387,000 tons of recoverable zinc and 101,000 tons of recoverable lead, were produced. The production had become delicately adjusted to price, which reached a cyclic peak for both metals in this year. Figure 20 shows the relation for zinc. With the increased economic competition in a market that had to adjust production to demand, there was an increasing tendency during the 1920's for the larger and more efficient mining companies to expand holdings and operations gradually at the expense of the small operators. Typically, the increased holdings were scattered in separated tracts throughout the field; many were held only temporarily and then abandoned for others.

In addition to the companies previously mentioned that had assumed an early lead in the development of the field, the Federal Mining & Smelting Co. increased its holdings, largely through a single deal in 1924, to where it ranked among the major producers, and was, indeed, the leading producer in the three years 1924–26. The St. Louis Smelting & Refining Co. also increased its operations in the early 1920's, to rate among the top producers, though its output was much lower in the latter half of the decade while it was developing large reserves near Baxter Springs. Other companies that commonly rated among the top 10 producers at different times in this decade were the Vinegar Hill Zinc Co. and the Century Zinc Co. (both subsidiaries of Youngstown Sheet & Tube Co., though the Vinegar Hill Co. had other affiliations before 1923), Consolidated Lead & Zinc Co., Anna Beaver Mining Co., Interstate Zinc & Lead Co., and New Chicago Mines Corp. Several expansions were effected through merging of smaller companies, the most notable being that in 1926 between the Underwriters Land Co. and the Consolidated Lead & Zinc Co., retaining the latter's name. The new company, which was a subsidiary of Eagle Picher Lead Co., rated, along with its parent company, among the top five producers of the field for the next few years. The Interstate Zinc & Lead Co. also attained its prominence in the same year through a merger of four earlier companies; its richest property was the Woodchuck mine.

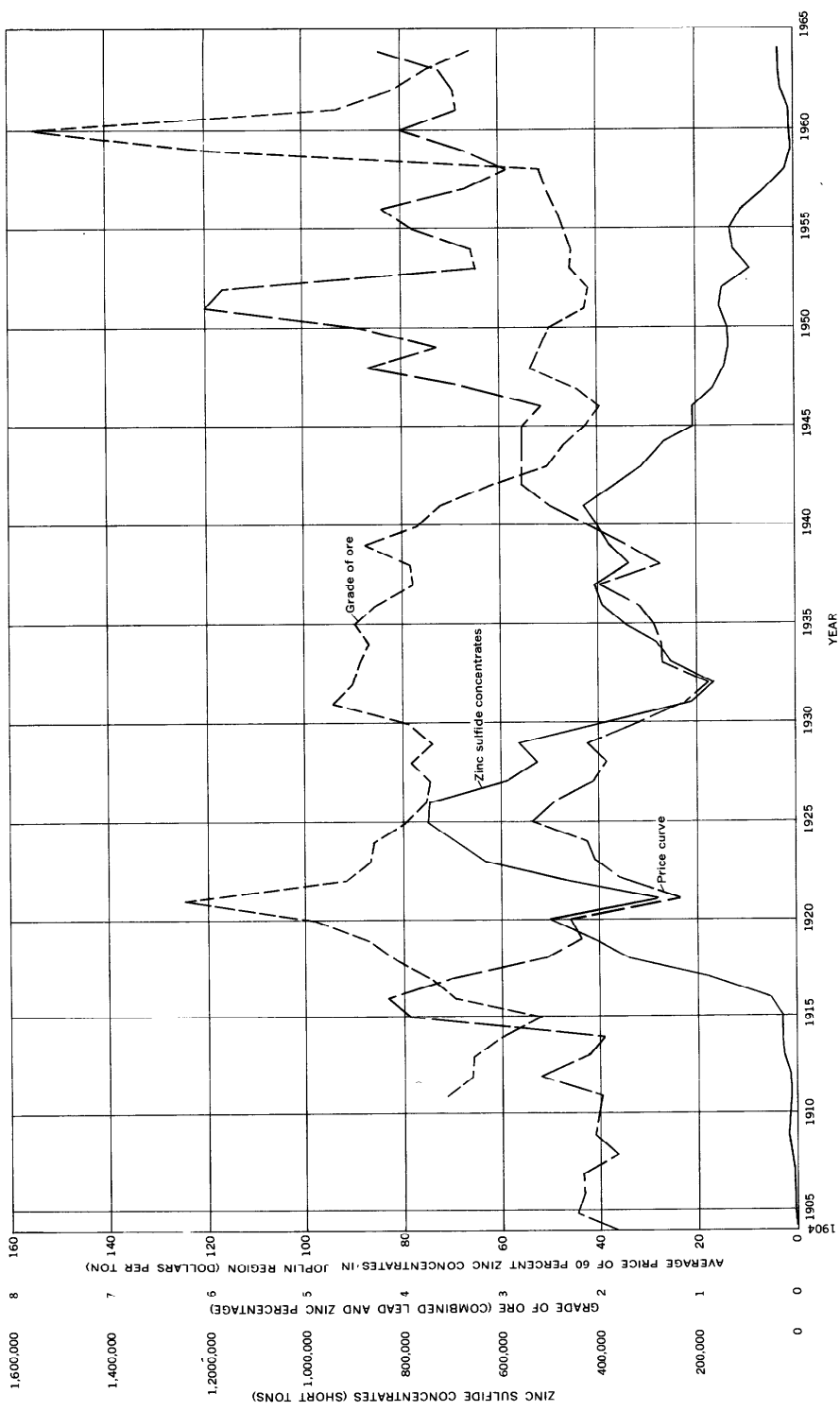

FIGURE 20.—Graph showing zinc sulfide concentrates production of Picher field and grade of ore in relation to selling price of standard 60 percent zinc concentrates. (The price paid to producers from 1942-47 was augmented over that shown in this chart by premium payments under the Premium Price Plan.) Average price from "Metal Statistics 1965," (American Metal Market, 1965, p. 504).

The Kansas Exploration Co., a subsidiary of St. Joseph Lead Co., began exploring a large tract in Kansas in 1921. The project failed to discover any extension of the field but resulted in opening a mine on the north edge. This company later acquired holdings in Oklahoma, though it never quite attained production that would rate it among the top half-dozen producers in the field. The American Zinc, Lead & Smelting Co. first became an operator in 1924 on a 20-acre tract; although it soon acquired other tracts, its large expansion in the field did not come until considerably later. Tri-State Zinc, Inc., subsidiary of New Consolidated Gold Fields, Ltd., began drilling of leased tracts in 1926 and bought a developed mine in the following year. However, this company attained its prominence in the field in later years through its extensive operation of tailings mills.

No sharp division can be made between these large companies and the many experienced and efficient companies of moderate size which also expanded or consolidated and continued to dominate the field in point of numbers. Mature development of the mining field had so increased the producing acreage that in each of the peak years, 1925 and 1926, respectively 184 and 193 different companies shipped concentrates. Nearly three-fourths of these companies exceeded $100,000 in gross annual income. In the 5 years from 1921 through 1925, the Picher field yielded 55 percent of the total zinc produced in the United States.

During the years of peak activity, a lead smelter was in operation at Hockerville, 3 miles east of Picher. It had been built in 1918 by the Ontario Smelting Co., and was sold in late 1923 to the Eagle Picher Lead Co. This company used it several years for the manufacture of antimonial lead. The plant was abandoned and dismantled in the early 1930's.

The 1920's marked the maturity of the field. Production generally declined in the latter half of the decade, but the underlying cause was the sensitive adjustment to price, rather than depletion; in any year a greater output could have been made at the prevailing price if activities had not been curtailed to avoid overproduction. Nevertheless, a few mines were worked out by 1927, and the number of exhausted mines increased steadily from that time.

The latter half of the 1920's also witnessed the ascendence of the tailings mill as an important factor in total zinc production. Although recovery of zinc concentrates from rerun of tailings had been started as early as 1909 in the part of the field around Commerce, and had been carried on to some extent through the field as it was developed in all following years, the introduction of the flotation process in several of the mills during World War I had been followed by general adoption of this process in the mid-1920's, and remilling of the huge tailings piles that had accumulated when only jigging and tabling had been used was a next logical step. Figure 21 shows the percentage of the total zinc production of the field that has been recovered from tailings. In many operations only zinc was recovered, but some tailings mills also recovered a little lead. In general, as a far greater percentage of the lead than of the zinc had been recovered in the initial milling, little lead was left in the tailings. Use of the flotation process as an adjunct of jigging and tabling in the last half of the 1920's and in later years insured the recovery of 80–85 percent of the metal contained in the crude ore, compared to the 58–70 percent recovery estimated for the older milling. The amount of zinc recovered in the concentrates from the milling of old tailings varied in different years from 1.28 to 0.19 percent of the tailings tonnage treated, expressed as an annual average for the whole field. In the early years (through 1936) when most of the tailings were being retreated for the first time, the annual average grade was never lower than 0.71 percent. Like the grade of crude ore, the

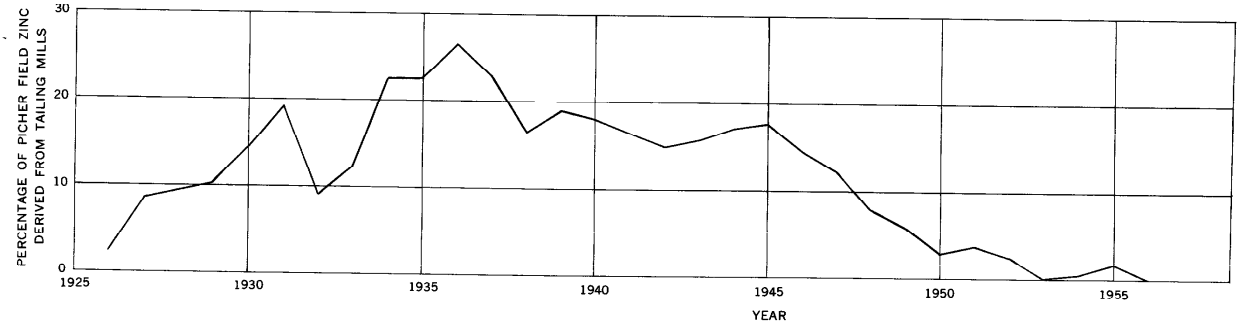

FIGURE 21.—Graph showing percentage of zinc production from reworking of tailings, Picher field (based on unpublished statistical data furnished by the U.S. Bureau of Mines). No production from tailings 1956–64.

grade of the tailings milled varied inversely with the price of zinc sulfide concentrates.

The zinc industry shared the economic doldrums of nearly all other industries during the depression of the early 1930's. Annual production dropped to a low of 168,000 tons of zinc concentrates in 1932, which was lower than any year since 1916. Pumping was discontinued in many mines, and the ground-water table, which had been kept below the level of mining by widespread pumping at many stations, was allowed to rise and partly fill some of the lower mine workings. By 1934, however, many of the mines had resumed operation, though with periods of inactivity to avoid overproduction, and most of the workings were pumped out. After the depth of the depression, the field showed a steady rise in production to 404,000 tons of zinc concentrates in 1937, and 430,000 tons in 1941, but this was only slightly more than half of the annual production for 1925. More and more mines were exhausted during the decade or were reduced to recovering the leaner blocks of ground, to cleaning up spots left in the earlier mining, or to salvaging pillars.

The 1930's also witnessed the growth of central milling in the field. The first mill built to treat ore from several tracts was the Bird Dog mill of the Commerce Mining & Royalty Co., completed in 1930. This plant was designed for operation at 2,750 tons capacity on a 24-hour basis, rather than the 10-hour basis common to mills up to that time (Isern, 1931). Haulage was by rail directly from the hoppers of the contributing mines. The success of this mill in sampling and milling ores from several different tracts indicated large economies in central milling over the practice that had prevailed up to that time (largely at the landowners' and royalty owners' insistence) of having separate mills on each 40- or 80-acre lease to insure proper royalty distribution. In 1932 the Eagle Picher Mining & Smelting Co., recently organized at that time to handle all mining-through-smelting activities of Eagle Picher Lead Co., completed a central mill near the southwest corner of the field, initially rated at 3,600 tons capacity but shortly thereafter stepped up to 5,500 tons. Other companies soon followed the example of hauling crude ore or tailings to a central company mill, though no other mills attained the size of operation of the Bird Dog and Eagle Picher mills, whose capacities were increased later in the decade, the latter to 10,000 tons. Many companies even abandoned milling altogether, and had their milling done on a custom basis by Eagle Picher or other companies. This procedure allowed more complete clean-up of depleted mines whose diminishing output was not sufficient to justify a milling operation. The central mills were particularly efficient in recovering the values from the more refractory siliceous ores in which the sphalerite is finely disseminated; hence, a larger proportion of their product was a flotation concentrate.

In this period of lowering ore grade, rising costs, and trend toward centralized milling, elimination of the small operators was accelerated. Among the larger operators that expanded in relative importance by acquisition of other holdings, the Eagle Picher Mining & Smelting Co. was by far the leader. Early in 1931 this company took over by foreclosure the holdings of Canam Metals Corp., which had acquired enough producing properties 2 years earlier to make it for a short time the seventh largest producer in the field, only to succumb soon thereafter, a victim of the depression. Later in the same year, Eagle Picher absorbed the holdings of its rich subsidiary, the Consolidated Lead & Zinc Co.; and in 1937 it took over the assets of the Mary M. Mining Co., which in the meantime had grown to be fourth largest producer, in part through purchase of the Admiralty Zinc Co. and Black Eagle Mining Co. holdings in the preceding year. In 1938 the Eagle Picher Co. eclipsed its earlier accessions by purchasing the entire holdings and assets of the Commerce Mining & Royalty Co., which was at that time the next largest operator in the field and held large ore reserves. The technical and operating staffs of the two organizations were integrated to great advantage in the merger. In 1940 the company obtained, through an intermediary, the Barr mine, which had been the most productive property of the Vinegar Hill Zinc Co. Other acquisitions were made during the decade that, individually, were not so large as those mentioned, but in the aggregate amounted to a sizable expansion.

Most of the other companies previously mentioned that had been among the leaders in the 1920's remained strong and active during all or most of the decade between the depression and World War II, except that the Anna Beaver Mining Co. was sold in 1929 to the Commerce Mining & Royalty Co. In addition, the Rialto Mining Corp., an old producer in the field, became relatively more prominent in this period of falling production rates, even though its maximum annual output had been made during the preceding decade. Midcontinent Lead & Zinc Co. and the Davis-Big Chief Mining Co., both of which shipped their ore to the Eagle Picher central mill, were also among the leading producers.

An outstanding feature of the decade was the relatively large production from tailings. Of companies operating largely or entirely on tailings, Tri-State Zinc, Inc., Cardin Mining & Milling Co., Semple Mining Co., and Captain Milling Co., each was among the top 10

producers of the field for one or more of the years from 1936 to 1941. Many other companies also reworked tailings, including some of the largest that operated primarily on crude ore. The peak of production from tailings came in 1936 (fig. 21) when 104,500 tons of zinc sulfide concentrates, milled from this source, contained 26.5 percent of the zinc produced that year from the whole field.

The period of World War II was marked by a steady decline in rate of production, in spite of a nationwide system of subsidized premium prices for base metals that was designed to pay a profit to each marginal operator above his production costs. Perhaps the immediate cause of the decline was the general labor shortage, which not only hampered direct production but also hampered exploration and development that are the necessary preclude to production. The developed reserves were gradually depleted without new reserves being developed. In spite of continuation of the Premium Price Plan to mid-1947, the lag in new development prevented any immediate reversal in the downward trend of production. The average metal prices toward the end of the decade were in general high, but the fluctuations and uncertainties, particularly in zinc price, were not conducive to any pronounced upsurge in production rate. The zinc price drop in 1949 produced a new low in output that year, lower than any year since 1916.

Although the decline in production rate during World War II had immediate causes in labor and material shortages, these merely delayed the eventual decline brought on by progressive depletion. During the life of the field, new ore was constantly being found and developed by drilling, but this was mostly within the boundaries of the field as fairly well defined by the early 1920's. An extension of the main field in the Melrose, Kans., area was discovered by drilling in 1925, but it proved to be small, and was quickly mined out during 1944–49. In the early 1920's, a new source had been recognized in the low-grade sheet-ground deposits at a lower stratigraphic level (Grand Falls Chert Member of the Boone Formation) in the northeastern part of the field, near Baxter Springs. These deposits were first worked in the Hartley mine, and although similar deposits were mined in the same general area in most succeeding years, it was not until the late 1930's and especially the 1940's that the great surge of sheet-ground production was made. The main leader in this development was the St. Louis Smelting & Refining Co. which worked several tracts simultaneously, all tributary to their Ballard central mill. At the same general time, a second large sheet-ground area was worked by Eagle Picher in the southwestern part of the field. Mining of these deeper deposits provided a temporary brake on the depletion rate, but they were not large in comparison to the original extent of mineralized ground in the field. The lean ore that had been passed up earlier was mined wherever it became economic, and conditions during the decade were especially favorable for this thus, the average grade of ore for 1946 was only 1.99 percent of combined zinc and lead as recovered in the concentrates (table 5).

A technologic development that contributed greatly to the economic recovery of progressively lower grade ores was the increasing mechanization of mining during and following World War II. The manpower shortage that had arisen made mechanization a necessity, but it proved to be a blessing in disguise, for the resultant economies delayed still further the final exhaustion of the field. Introduction of slushers in sheet-ground mines in the late 1930's and of track-mounted shovels in the early part of the war largely did away with the traditional method of hand shoveling the ore into steel cans at the working face. Rubber-tired diesel trucks of 10-ton capacity were perfected for underground haulage, starting in 1946, and within 3 years, 25 of these trucks were in steady operation by the Eagle Picher Co., in combination with diesel Caterpillar loading equipment (Clarke, 1949). Other companies quickly adopted the same methods. Trackless loaders and haulage give a greater flexibility to underground mining during cleanup operations in that scattered spots of ore can be quickly and economically mined in extensively cut ground, whereas the laying of track to each spot under the older system of electric train or tail-rope track haulage and track-mounted shovels would be impractical. Commonly, what appeared to be a small spot of ore would lead into an unsuspected block of considerable size that could not have been recovered under other conditions.

The tailings available for retreatment were likewise depleted. During the war the "primary" tailings were exhausted, and there was a definite transition to second run and even third run of some of the tailings. The lowest average annual recovery of 0.19 grade percent zinc from tailings was reached in 1946.

In the period since 1940, the Eagle Picher Co. has dominated in the production from the field, usually producing two to six times the output of its nearest competitor. Its central mill was expanded to a capacity of 15,000 tons per day, though part of this capacity is for custom ore. Other leading companies have included

the St. Louis Smelting & Refining Co. (after 1948 recorded as a division of the National Lead Co.), Federal Mining & Smelting Co., Davis-Big Chief Mining Co. (which in 1941 acquired the mines of Skelton Lead & Zinc Co.), Rialto Mining Corp., Evans Wallower Zinc, Inc., Tri-State Zinc, Inc. (tailings), and Cardin Mining & Milling Co. (tailings). After the Federal central mill (Gordon mill) burned in 1943, most of its ore was milled on a custom basis by the Eagle Picher Co. until the Federal company abandoned the field 10 years later. In 1948 the Nellie B. Mining Co., new in the field, acquired extensive holdings largely in a block east and southeast of Cardin, including the remaining assets of Evans Wallower Zinc, Inc., those of Rialto Mining Corp., and from the Marcia K. Mining Co., the properties that a few weeks earlier had belonged to the Davis-Big Chief Mining Co., as well as additional properties of the Marcia K. Mining Co. In 1951 the American Zinc Lead & Smelting Co. bought out the Nellie B. Mining Co. to become the second largest producer in the field.

Because of depressed metal markets, many operations were cut back or suspended in 1957, and by midyear of 1958, all the major mining operations were closed in the most complete shutdown of the field's history. The National Lead Co. dismantled and removed its central (Ballard) mill in 1959; its mining equipment, facilities, and most of its leases were acquired by the Eagle Picher Co.

Mining was resumed at a reduced rate in 1960 and has gradually increased, though (as of 1964) it has not reached the status that prevailed before the 1957 curtailment.

As of 1964, many properties have been worked out and abandoned, and others are reaching this stage one at a time. The bulk of the remaining reserves in the field are marginal in grade and can be mined only so long as economic conditions remain favorable. Because the Eagle Picher Co. has integrated activities in the zinc and lead industry that include smelting and refining as well as fabrication and marketing of final products, this company may be able to mine and mill lower grade ore than a company that makes its entire profit on the concentrates sold. Hence, it may be expected that this company will gradually acquire such properties as are abandoned by the other companies if any marginal reserves remain. Any economic recession that could lead to the abandonment of pumping in the field for any time, as in 1930, would probably result in loss of the remaining low-grade reserves, for the mine workings are so extensive and so interconnected that the cost of pumping them out again would be prohibitive when balanced against the tonnage and grade of the remaining reserves.

MINERALOGY OF THE ORES

Sphalerite and galena are the commercial ore minerals. They are accompanied by a little chalcopyrite and enargite but in amounts so small that the contained copper is not recovered. Marcasite and pyrite are common associates of the ore minerals, though the pyrite is usually in such small grains and so sparse that its effect in the ore is negligible. Wurtzite was not found in the Picher field but is present at one mine near Joplin. Gangue minerals include jasperoid, dolomite, calcite, and locally, a little quartz or barite. In the early days of mining in the Lincolnville area southeast of Quapaw, some smithsonite and calamine were marketed from surficial deposits, but these minerals are no longer of more than accidental occurrence in the field. The galena locally has alteration rims of anglesite, some of which, in the deeper mines of the western part of the field, formed after the ground was opened by mining. Sulfates of calcium, iron, zinc, and magnesium, also formed after mining began, are common as efflorescences in some of the mine workings but are insignificant in actual bulk. Although neither a primary nor secondary gangue material in the strict sense, a thick black petroleum or light tar that originally had been trapped on top of the ground water in the structural highs beneath the impervious shale cover has seeped down upon removal of the mine water and permeates the ore-bearing ground or floats on the underground sumps in many places (Fowler, 1933).

In its major features the mineralogy of the Picher field does not differ greatly from that prevailing in the rest of the Tri-State region. There may be some minor differences between Picher and Joplin in the crystal form assumed by sphalerite, which are mentioned below. The Joplin area contains at least one deposit which is unique, in comparison to other Tri-State deposits, in composition and form of its zinc sulfide. This deposit is in the Zig Zag mine which was being worked in the mid-1930's. To present a more nearly complete account of Tri-State mineralogy, and particularly since the type of mineralization may have genetic significance, the mineralogy of the Zig Zag mine is discussed in this report.

In the following discussion the minerals are divided into primary minerals and secondary minerals. In the primary mineral group, the sulfides are treated before the gangue minerals.

PRIMARY MINERALS

Sphalerite (*black jack, rosin jack, ruby jack, blende*—ZnS; Zn *67 percent*, S *33 percent*)

Sphalerite is the most abundant of the primary ore minerals. In typical replacement ore the sphalerite is massive and rather coarse grained, having cleavage faces that are commonly ⅛ to 1 inch across. Crystals of sphalerite in solution channels, vugs, and caves are almost equally abundant and characteristic. These crystals commonly range in diameter from a small fraction of an inch to 4 or 5 inches, and Weidman (1932, p. 53) reports crystals more than 1 foot in diameter. The smaller crystals are generally well formed, whereas the larger ones are usually imperfect and irregularly twinned. At the opposite extreme are the fine-grained disseminations, pinhead size or smaller, in jasperoid. Finely disseminated sphalerite is always crystalline but is usually anhedral or only partly euhedral against the jasperoid.

The crystal form of the sphalerite that develops free in openings varies with the color. Black jack, which is a type of sphalerite in dolomitic ground, crystallizes as a combination of dominant negative and subordinate positive tetrahedrons, accompanied by the cube (fig. 22). A very few crystals have, on the negative tetrahedron, poorly developed beveling faces of a trigonal tristetrahedron (m, $\bar{1}13$).[7] The positive tetrahedron may be lacking in the smaller crystals, or in some crystals as large as three-quarters of an inch in diameter; and rarely, the cube faces may also be lacking, leaving only the single tetrahedron. Well-formed crystals may be as much as 2 inches in diameter, and larger crystals are commonly defective only because of twinning. The crystals are unusualy black, opaque, and submetallic in appearance, but all have a rosin-colored core when broken, though considerably darker and duller than that of the other types of sphalerite. The black jack as here described was early recognized as a distinct type (type 1) by A. F. Rogers, who states (Rogers, 1904, p. 457) that it is perhaps the commonest of the types in the Galena-Joplin district.

In widely scattered localities the black jack may be in roughly hemispherical rosettes instead of well-formed crystals. Such rosettes commonly have an uneven warty surface made up of small poorly formed crystal units roughly shingled in parallel orientation. The core is rosin jack, which may have cleavage surfaces as much as 5 inches across.

[7] Crystal forms here cited have been checked by Marie L. Lindberg, U.S. Geological Survey. Reflections of the faces on the goniometer are very poor, but they give average angles that are closest to the common forms cited.

Rosin jack and ruby jack crystals usually have a combination of the negative tetrahedron with the corresponding negative trigonal tristetrahedron (m, $\bar{1}13$), modified in part by the cube (fig. 23). In different crystals at a given locality, either the tetrahedron or trigonal tristetrahedron may dominate to the exclusion of the other and to the exclusion of the cube, but all three forms are normally present if several crystals are examined. Thus, the essential crystallographic difference from the black jack lies in the tendency for the trigonal tristetrahedron to develop in all or most of the crystals from a given locality, and in the absence of the positive tetrahedron. Well-formed rosin or ruby jack crystals are usually a quarter of an inch or less in diameter, rarely as much as half an inch. In many localities, such

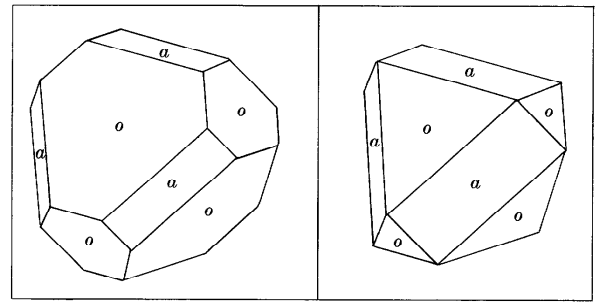

FIGURE 22.—Characteristic crystal forms of black jack (sphalerite) in Picher field. Drawings modified from Flink (1888) with different orientation, showing left-side cube face instead of conventional right side.

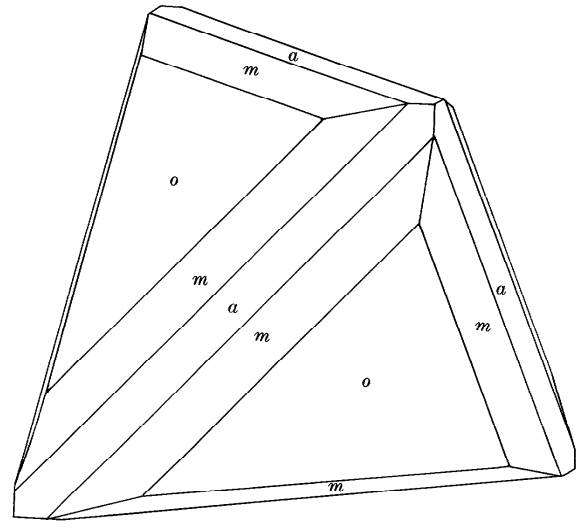

FIGURE 23.—Characteristic crystal form of rosin jack and ruby jack (sphalerite) in Picher field.

crystals may be variegated in color, the points (in 111 position) being darker than the centers. Larger crystals are characteristically rounded or irregular and poorly formed, although the influence of the crystallographic forms that characterize the smaller crystals can usually be recognized. The larger crystals also tend to be darker colored, though still translucent on the thinner edges.

The rounding of crystal forms in the rosin and ruby jack is probably due to the interfering influence of the dodecahedron in competition with the trigonal tristetrahedron. Much of the rounding in small and medium-sized crystals is in such a direction, but no clearcut dodecahedral face has been recognized in the crystals from Picher. In other zinc districts the dodecahedron is a dominant form, and Gebhardt (1933, p. 34–36) describes several yellow to ruby-red crystals from Joplin that show dodecahedral faces. These crystals, however, are all rounded. Rogers (1904, p. 453–58) describes two types of crystals from and near Galena, Kans., in which the dodecahedron is the dominant form. It is possible that at the time of ore deposition, the physical and chemical environment at Galena and Joplin may have differed somewhat, at least locally, from that at Picher, even though the black jack type of sphalerite, with its characteristic crystal form, is common to the two areas.

Quantitative spectrographic analyses have been made of seven sphalerite samples by Janet D. Fletcher, of the U.S. Geological Survey. Six of these are in pairs, representing the differently colored peripheral and core blocks, respectively, of nonhomogeneous crystals. The results are given in table 6. As a check, M. K. Carron, of the U.S. Geological Survey, re-analyzed one of the samples chemically (No. 7) for the more abundant metallic constituents, using standard analytical procedures, with the results shown at the end of the table.

The first four samples, from two crystals collected 2,000 feet apart, have consistent differences in the distribution of several elements between black jack shells and rosin jack cores: the black jack shells contain more iron, copper, lead, titanium, manganese, and silver and less cadmium, gallium, and cobalt. Other elements are not significantly partitioned between different parts of the crystals, though there is a suggestion that indium is concentrated in the shells and barium in the cores.

Samples 5 and 7, representing the ruby jack phase, are alike in having the lowest iron content and the highest copper, germanium, indium, and perhaps, also, gallium contents. The composition is most like the rosin jack phase when certain other elements are considered—namely, lead, manganese, cobalt, and silver. Cadmium content is not consistent as between the two samples, but is lower than in the rosin jack and roughly comparable to the black jack. Other elements in the ruby jack analyses have no consistent correlation with the type of sphalerite, either because of a spread in the ruby jack analyses or a lack of partition between the different color types.

TABLE 6.—*Spectrographic analyses of sphalerite, in percent* [1]

Sample	Fe	Cd	Cu	Ge	Ga	In	Mg	Pb	Ba	Ca	Ti	Mn	Co	Ag
1	0.29	0.60	0.088	0.02	0.003	0.01	0.02	0.02	0.003	0.003	0.003	0.001	0.000	0.0004
2	.14	.88	.052	.02	.044	.01	.02	.001	.01	.002	.000	.0007	.0009	.0000
3	.31	.29	.084	.03	.003	.02	.02	.02	.003	.002	.002	.002	.000	.0003
4	.17	.88	.044	.02	.040	.01	.007	.001	.005	.003	.000	.0006	.0009	.0000
5	.12	.72	.12	.1	.044	.06	.01	.004	.004	.004	.003	.0006	.0009	.0000
6	.15	1.4	.016	.04	.003	.0006	.01	.000	.003	.004	.003	.0007	.0008	.0000
7	.11	.31	.18	.1	.060	.1	.04	.003	.002	.003	.000	.0005	.0009	.0000

Standard chemical analysis of sphalerite, in percent

Sample	Fe	Cd	Cu	Ge	Zn
7	0.15	0.30	0.18	0.06	66.36

[1] Looked for, but not found: Au, Hg, Ir, Pt, Mo, W, Sn, As, Sb, Bi, Tl, Ni, Cr, V, Sc, Y, Yb, La, Zr, Th, Nb, Ta, U, Be, Sr, P, B. For other tests for mercury, see p. 158.

1. Lawyers mine, NE¼NW¼ sec. 14 (a partial section), T. 35 S., R. 23 E., Cherokee County, Kans.; 180 ft west of east line, 400 ft north of south line (=State line). Black jack shell from a crystal 1½ inches across. This shell on much of crystal is about ⅛ in. thick, but on the 111 face (positive tetrahedron) it thickens as a pie-shaped wedge pointing toward center of crystal. Sample is about 90 percent from outer ¼ in. of this 111 face; the remainder is from an adjacent cube face.
2. Rosin jack core from same crystal as sample 1.
3. Midcontinent mine, SE¼SW¼ sec. 11, T. 35 S., R. 23 E., Cherokee County, Kans.; 330 ft east of west line, 1,090 ft north of south line. Black jack shell from a large compound crystal 3 in. long, 2 in. wide, and 1½ in. thick, all parts of crystal having same orientation.
4. Rosin jack core from same crystal as sample 3.
5. Fox mine, SE¼NE¼ sec. 11, T. 35 S., R. 23 E. Cherokee County, Kans.; 10 ft south of north line, 65 ft west of east line. Ruby jack from two rounded and poorly formed crystals, growing 2 in. apart, each about ½ in. in diameter. About half of each crystal taken, the rest being rejected because of certain pale or colorless splotches in the interiors of the crystals.
6. Colorless or very pale blocks in interior parts of the two crystals from which sample 5 was taken.
7. Anna Beaver mine, NE¼NE¼ sec. 19, T. 29 N., R. 23 E., Ottawa County, Okla.; 780 ft west of east line, 1,050 ft south of north line. Many small crystals of ruby jack, average 1/16 in. diameter, picked from finely crystalline jasperoid surface in vug.

Sample 6 represents a phase of common occurrence inside crystals of other types, though it has no great importance in actual mass. The analysed sample suggests that this nearly colorless material may carry the highest percentage of cadmium. It is not so low in iron as might have been expected, in comparison to the other color types.

The copper found in these specimens is of special interest, inasmuch as that reported in other analyses of sphalerite is commonly presumed to be present as microscopic inclusions of chalcopyrite rather than in the sphalerite lattice. The two crystals whose rims and cores supplied the first four samples are alike in containing a little fine crystalline chalcopyrite, easily visible under × 10 hand lens, imbedded at the contact between the black jack shells and rosin jack cores. This contact material was carefully avoided in preparing the samples. All fragments entering into the samples were examined under the hand lens to insure homogeneity; and crystals adjacent to those sampled, when examined in thin and polished sections, contain no recognizable chalcopyrite except in the rejected narrow contact zone. Furthermore, it seems significant that the ruby jack, which has no visible chalcopyrite in thin section, has the highest percentage of copper associated with the lowest percentage of iron. In the spectrographic analysis of sample 7, the copper is even higher relative to total iron in the sample than the stochiometric ratio for chalcopyrite, although the discrepancy is virtually eliminated in the standard analysis of this sample, and there is a slight stochiometric excess of iron in the ruby jack of sample 5. As a certain amount of Fe is to be expected in the ZnS lattice, there is little reason to assume that the minimal amount shown in these rosin jack specimens is in an invisible chalcopyrite constituent.

This conclusion is in accord with results published by Barrett (1940), who found 0.015–0.08 percent copper in specimens of sphalerite from the Picher area, carefully selected under the microscope to avoid inclusions of any kind.

Spectrographic investigation of minor elements in banded schalenblende from the Aachen-Moresnet region along the German-Belgian border shows that iron, lead, manganese, and germanium are concentrated in the dark bands and cadmium in the light bands (Kutina, 1953). Silver has no recognizable preference for either band. Studies of similar material from Poland, although not so decisive, nevertheless show a concentration of iron, lead, and germanium in the dark bands. Although this European material is believed by Kutina to have been deposited colloidally, there is a rough analogy to the partition of certain elements as between the black jack shell and rosin jack core of crystalline material from the Picher field (table 6, samples 1–4). Similar partition is shown by the iron, lead, cadmium, and manganese. On the other hand, the silver, which is partial to the dark material in the Picher ores, has no preference in the European ores; whereas germanium, partial to the dark material in the European ores, has no significant differentiation in the analogous Picher examples, though it has a decided preference for the Picher ruby jack. Kutina concludes that the iron, cadmium, manganese, and germanium (as well as thallium and mercury, not found in the Picher material) are isomorphous with the zinc, whereas other constituents, but notably the lead, are present as heterogeneous impurities.

Thin sections of black jack crystals adjacent to one of those analysed from the Picher field show that most of the coloring matter of the shell is very heterogeneously concentrated in splotches, indefinite concentric bands, and radial wisps, making up perhaps 30 percent of the field. This pigment is mostly transparent and is dark purple. The color is best seen when the condensing lense of the microscope is used, but it grades in certain narrow lines to near opacity. Polished sections under highest magnification reveal nothing but sphalerite, and powder X-ray pictures made by Fred A. Hildebrand of the Geological Survey, in search of other materials, likewise yielded only the sphalerite pattern. It is doubtful, however, that the very small percentages of copper and lead revealed by the spectrograph would be distinguishable in the X-ray picture even if they were present in heterogeneous impurities. Other constituents, particularly those isomorphous with zinc in the sphalerite lattice, might well be present in the clear groundmass of the black jack shell, which is somewhat grayer than the groundmass of the rosin and ruby jack.

Cross sections of black jack crystals show that the black phase shell is thicker on the positive tetrahedron face (111) than on the cube or negative tetrahedron faces. The boundaries form an irregular wedge pointing toward, though not extending to, the center of the crystal. Thin sections also show a denser concentration of the dark coloring material in this wedge. When it is remembered that the positive tetrahedron face is characteristic of the black jack and missing in other color phases, it would appear that there is a genetic connection between the 111 face and some constitutent or constituents responsible for the black color. As mentioned on page 103, variegated crystals of rosin jack or ruby jack have the dark color on the points built out

in the 111 direction, although the 111 face is not developed in such crystals.

The rosin jack cores of black jack crystals have, in thin section, irregular but roughly concentric banding of bright orange yellow in a nearly colorless sphalerite base. The yellow phase makes up roughly 40 percent of the core, with the most intensely colored and widest band at the outside, next to the black jack shell. To avoid the inclusions of chalcopyrite which lay on the border between the core and the shell, this border zone was rejected in preparing the sample of the core for analysis. Hence, the analysed sample contains less of the yellow phase than a strictly representative sample. Thin sections of the ruby jack phase show an even more intense concentration of the orange-yellow coloring, chiefly in a peripheral zone but also in irregular blotches in the colorless crystal core and along certain directions parallel to the cleavage.

Most of the zinc sulfide at the Zig Zag mine near Joplin is dark brown and stalactitic, and has no external crystal form. Optical examination shows that part of it is wurtzite, but a larger amount intermingled with the wurtzite is isotropic and presumably sphalerite. Marcasite-lined vugs in the stalactitic material hold a few scattered crystals of sphalerite, 1–2 millimeters across, with rounded and undeterminate crystal faces, and of a unique pale-golden color, resembling in this respect the golden calcite of the district. This sphalerite is so unlike other Tri-State sphalerite as to suggest that it may be a secondary phase, related to the present surface of weathering.

Tri-State sphalerite or blende concentrates, as a commercial commodity, are sold at a quoted price per ton for the standard grade containing 60 percent zinc. Any excess of zinc over this amount receives a premium, and any deficiency, a discount from the quoted price. Penalties have been assessed for impurities, such as iron (marcasite) and lead, the amount varying at different times in the history of the Picher field, and flotation concentrates have at times been penalized in comparison to jig and table concentrates of comparable grade.

Although the average grade for many years has been close to the standard 60 percent, many shipments have rated above this. Some of the richest concentrate on record is that shipped by the Quebec mine in October 1918, when five cars are reported to have run 64.8, 64.9, 64.9, 65.0 and 65.7, or an average of 65.0 percent zinc for the total shipment. The Bendelari mine shipped a carload of flotation concentrate in 1931, averaging 65.4 percent zinc.

For eight of the years from 1921 to 1930, "Mineral Resources of the United States" (U.S. Geol. Survey, 1921; U.S. Bur. Mines, 1924–30) reported the average content of cadmium in the marketed concentrates of the Picher field, based on the many settlement assays for the different shipments. The amount ranged between 0.37 and 0.56 percent. As the cadmium content can be related only to such variables as the type of sphalerite, and conceivably, though not demonstrated, to geographic source of the sphalerite within the field,—none of which are affected by such economic factors as the grade of ore profitably minable or the proportion of flotation concentrates in the total output,—it is believed that the limits of cadmium content quoted above should be valid throughout the total history of the field.

Wurtzite (ZnS; Zn *67 percent*, S *33 percent*)

The zinc sulfide at the Zig Zag mine in north Joplin (NW¼ SW¼ sec. 35, T. 28N., R. 33W.) is dark brown and stalactitic, the individual columns ranging from a small fraction of an inch (one sixty-fourth inch) to perhaps 1 inch in diameter. Usually, many stalactites are fused together to a compound mass that may be several inches in diameter. The outer surface of such a mass shows the fluting of the individual columns, and the cross section shows the irregularly polygonal ends of the columns separted by thin interstitial voids or by massive FeS_2. Some of the stalactites separated by interstitial marcasite may be round, but usually the cross section is only irregularly rounded. The columns show a rough radial crystallinity that may be accentuated in places by radial blades of galena. The center of radiation, which may be eccentric in some of the stalactites, is commonly marked by a fine hollow capillary tube parallel to the axis of the stalactite. In places, the form may be mammilary or botryoidal. Thin sections and crushed fragments show that the major part of the zinc sulfide is isotropic and hence sphalerite; the rest is anisotropic and possesses the other optical properties of wurtzite.

Thin sections across the stalacites show an irregular concentric banding in different shades of brown and of varying thickness; some bands are well defined and others vague. Contacts between adjacent bands are commonly marked by a dark-brown line. The banding is generally scalloped, and in some bands the scallops may bow inward toward the central capillary opening (fig. 24). There are anastomosing fine dark lines which follow no recognizable pattern in peripheral parts of the stalactite, but which organize into an imperfect pattern of complementary radial and concentric lines near the central capillary opening.

Under crossed nicols, from 5 to 20 percent of the material, varying in different bands, has the anisotropism of wurtzite. It is in ill-defined, very fine grained units, which include many fine fibrous-appearing units

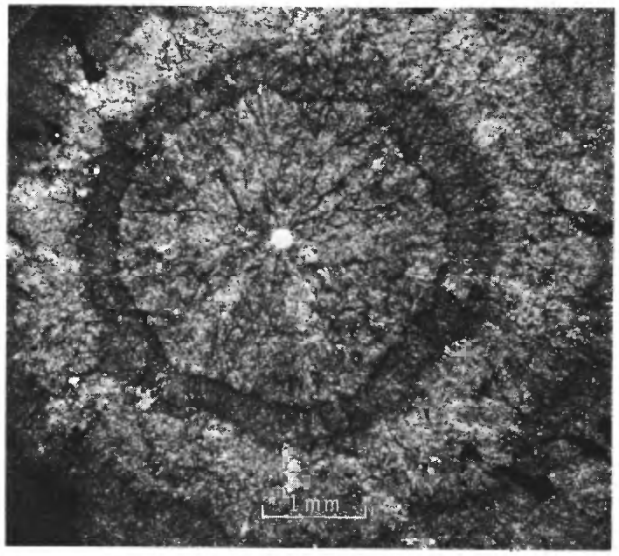

FIGURE 24.—Photomicrograph of thin section across stalactite of mixed sphalerite and wurtzite from Zig Zag mine, showing concentric banding, and roughly radial dark lines near the central capillary opening. Plain light.

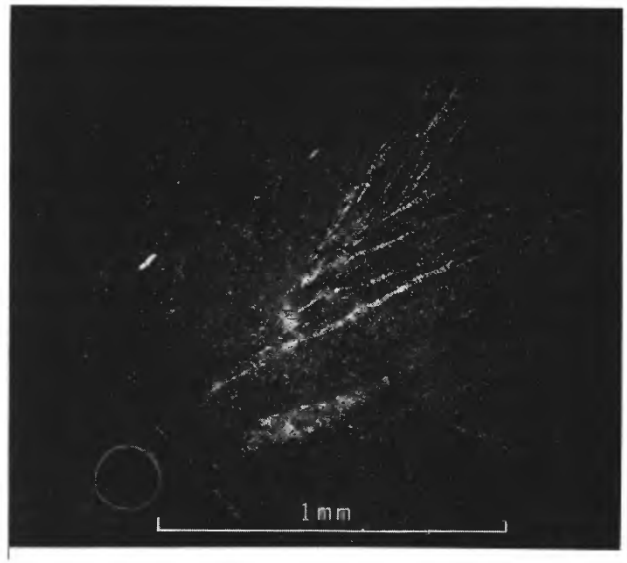

FIGURE 25.—Photomicrograph of thin section across stalactite of mixed sphalerite and wurtzite from Zig Zag mine. Light streaks are wurtzite. Capillary opening at lower left. Crossed nicols.

as seen in cross section. These are usually oriented radially, but they are also arranged like the barbs of a wheat head in relation to a concealed shaft that is oriented radially (fig. 25). What appear to be fibers are, without much question, cross sections of platy wurtzite crystals flattened parallel to the basal pinacoid. Their optical elongation, as seen in thin section, is negative.

As the "fibrous" crystals within a band always terminate at the outward border of the band, it is believed that they were formed during the growth of the stalactite, those in each band forming before the next layer was deposited. They now are phantom crystals with indefinite boundaries, discontinuous, and mottled with isotropic material in minute detail, and are undoubtedly partly inverted to sphalerite.

A small percentage (estimated 5 percent) of wurtzite was confirmed by X-ray pictures taken of the powdered material by Brian J. Skinner, of the U.S. Geological Survey; this was the 2H polytype in the terminology used by Frondel and Palache (1950). The original material probably contained a higher percentage.

Most of the stalactite surfaces are rough and noncrystalline in detail. Some, however, bristle with small chestnut-brown inverted horns that are smaller at the attached base than at the upper free end. The general shape is that of a minute prickly pear fruit. A few of the more perfect ones, however, reveal a hexagonal free end, and are obviously steep hexagonal pyramids with a basal pinacoid on the free end, attached to the stalactite by the smaller end that would otherwise have formed the peak of the pyramid. Most crystals average about 0.5 mm in diameter on their free base and 1 mm in height, though a few are as much as 1.5 mm long with a correspondingly thicker diameter. A few crystals have a stubbier pyramidal shape. The crystal form is that of wurtzite; crystallographic and X-ray diffraction studies show that the steeper pyramids represent a new polytype, wurtzite 10H, and the stubbier crystals represent a new occurrence of the polytype wurtzite 6H (Evans and McKnight, 1959). Some crushed crystals have a fine polysynthetic banding of isotropic sphalerite parallel to the base. Some of the wurtzite crystals have a second, and usually smaller, crystal growing from their free ends, in parallel orientation.

Most crystals have a scaly layering developed parallel to the base that is accompanied by a ragged surface and a change of color to dull honey yellow, though most crystals still have a dark spot in the center of the base. These changes appear to be an alteration effect, but whether they existed at the time the specimens were collected or have developed over several years in the laboratory is not known. The final product resembles the pulverulent pinkish-drab alteration product which is formed from the stalactitic material where it is adjacent to oxidizing marcasite. Possibly, the microscopic banding of sphalerite parallel to the base of wurtzite

crystals as previously mentioned may be an incipient phase of inversion to sphalerite.

An occurrence of wurtzite crystals associated with stalactitic massive material believed also to be wurtzite has been described by Rogers (1904) from a locality in east Joplin, about 1½ miles southeast of the Zig Zag mine. Siebenthal (1915, p. 258–62) surmised that this locality was at the Combination mine, and has given further descriptions of the material that show the close similarity of the mineral occurrence to that at the Zig Zag mine.

Galena (*"lead"*—PbS; Pb *86.6 percent*, S *13.4 percent*)

Although not so abundant as sphalerite, galena is, nevertheless, widely distributed; in only a few mines is it too sparse for economic recovery. It occurs disseminated in jasperoid or dolomite (gray spar), with other minerals in replacement seams and fractures along and across the bedding, and as crystals in vugs, solution channels, and caves.

Galena crystals disseminated in jasperoid may be anhedral, but they are more commonly euhedral, the overall shape being somewhat irregular. Where in contact with disseminated sphalerite, the mutual boundaries are very irregular, and there may be minute rounded, elongated, or wedge-shaped inclusions of the galena within the sphalerite along and near the contact. Similar inclusions of galena occur within the margins of pink spar dolomite masses where the two minerals are interbanded in replacement ore. In these latter inclusions the tendency for the galena boundaries to follow rhombic cleavage directions in the dolomite suggests that the galena has replaced the dolomite.

Crystal sizes average somewhat coarser than in the associated sphalerite, though maximum crystal sizes are comparable in the two minerals. Disseminated crystals in jasperoid are usually ¼ to 1 inch across, but in places may be as much as 6 inches across. Typical crystals in open ground are cubes, in part with octahedral truncation of the corners, and range usually from ¼ inch on an edge to 2 or 3 inches, less commonly to 5 or 6 inches. The largest crystal noted was an 8-inch cube on the Kenoyer tract (NE¼SW¼ sec. 20).

Octahedral crystals are much less abundant and are of local occurrence. The octahedrons are smaller than the average cube crystals; a maximum length of 1½ inches is only rarely found, and most octahedrons are less than 1 inch in length. We have been unable to correlate the ocurrence of the octahedral form with other geologic features, though, in general, the octahedron is later than the cube in the paragenetic sequence. A common feature is the presence of small octahedral pyramids on the surfaces of galena cubes, but particularly on the corners and edges. Those on the cube face may be elongated into a diagonal ridge by overextension of two of the octahedron faces. At one locality in the Barbara J mine, octahedral pits occur on a cube surface.

In places (Tulsa Quapaw and Robinson mines), the form disseminated in jasperoid is the octahedron, whereas the form in closely adjacent vugs is the cube. But elsewhere, all intergrades from octahedron to cube have been seen in crystals disseminated in jasperoid.

At one locality in the southwestern part of the Goodwin mine, the galena crystals have a peculiar platy habit due to abbreviation in one of the 3 cubic dimensions. These platy crystals have octahedrally truncated corners. A typical crystal measured about ⅜ inch square and only 3/64 inch thick. Although formed in an open vug, the crystals are also unusual in that several are warped over the dolomite crystals on which they rest, simulating a flexibility that does not exist. The explanation for this phenomenon is not evident, but if at an early stage in the crystal growth the plate was very thin, it might have been warped or fractured by disproportionate growth on the distal edges, and such warping would have been "frozen" by later additions to the crystal.

In a part of the Brugger mine, the crystals are a combination of octahedron and cube but are elongated parallel to a dodecahedral axis, yielding a puzzling crystal that simulates a hexagonal prism surmounted by four faces in a pyramid.

Many galena crystals, regardless of crystal form, contain internal voids. These may be lined by crystallographic faces or may have smooth irregular walls. Some are obviously connected with the outside and may carry small late crystals of sphalerite or chalcopyrite; others may originally have been sealed off, though no liquid has been found in any of them.

In the Zig Zag mine, galena occurs imbedded in the stalactitic sphalerite and wurtzite as small irregular segregations, but also as roughly radial thin blades, terminating outward on the stalactite surface in small pyramids of octahedral crystallization, usually with cube modification on the points. Polished sections show that the galena blades embedded in some of the smaller stalactites may be microscopic in size, 0.01–0.001 mm, or even smaller in cross-sectional length. Some of the larger stalactites show on the broken cross section a reticulated network of fine curving galena blades and less regular masses in parallel orientation over distances of 2 inches, as shown by the cleavage. The external crystal terminations may also be in parallel orientation.

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A pronounced etching of the galena in many places, amounting in extreme cases to complete resolution, and with variable regrowth of the galena, are more appropriately described in a general discussion of etching phenomena in relation to several of the sulfides (p. 125).

According to Barrett (1940), the galena from the Picher field carries 0.1 ounce silver per ton, 0.006 percent copper, and 0.003 percent iron.

Tri-State galena concentrates are sold at a quoted price per ton for the standard grade containing 80 percent lead content, with a premium for any higher grade and with a discount for any lower grade. The marketing system is similar to that for sphalerite.

Chalcopyrite ("*copper*"—$CuFeS_2$, Cu *34.5 percent*, Fe *30.5 percent*, S *35 percent*).

Chalcopyrite is widely distributed, but nowhere in great enough abundance to constitute a source of copper. It occurs mainly in vugs and caves, and is practically confined to ground containing pink spar dolomite, though it may occur more abundantly on crystalline sphalerite that is present along with this dolomite, or, rarely, on galena. Dolomite is sparse in the southwestern part of the field and so is chalcopyrite.

Chalcopyrite crystallizes in sphenoids (nearly tetrahedral in aspect) or modified sphenoids that are usually about a quarter of an inch on an edge, but the largest crystals, which always grow on pink spar dolomite, are as much as three-quarters of an inch on an edge. Crystals growing on a sphalerite crystal are commonly in parallel orientation, so that their faces flash in unison as the light is reflected from them. Exceptionally, such crystals may be abundant enough to form a nearly continuous coating of the sphalerite. Chalcopyrite that is early in the paragenetic sequence may underlie crystals of sphalerite or galena that are later deposited in the vug or cave.

Chalcopyrite also occurs, though only in small amount, as minute inclusions in sphalerite. Such inclusions may be irregularly rounded, or on the other hand, they commonly have straight-line crystal boundaries on one or more sides. They are either irregularly and sparsely scattered, or arranged zonally in the sphalerite crystals, more abundantly toward the surface of such crystals. They thus show gradations to deeply imbedded overgrowths on the sphalerite. As the straight-line crystal boundaries are usually faced toward the surface of the sphalerite crystal, these inclusions by and large must represent small euhedral crystals deposited at a stage during the growth of the sphalerite crystal. Inclusions that lie in zonal lines evidently record short pauses in the growth of the sphalerite crystal. There is no resemblance to textures that have been ascribed to exsolution in the solid state, or to replacement (Bastin, 1950). Rarely, blebs of chalcopyrite have been found in galena crystals near the surface, associated with better formed chalcopyrite perched on top of such galena crystals. The mode of emplacement is undoubtedly the same as that in the sphalerite.

Chalcopyrite has replaced to some extent the pink spar dolomite on which it commonly grows. It may invade along a cleavage crack of the dolomite, or may form irregular blebs in the dolomite with a marked tendency for stretches of the boundary to lie parallel to a cleavage direction or a cleavage angle of the dolomite crystal. Such blebs may contain irregular islands of the replaced dolomite. The replacement boundary may be minutely jagged (teeth 0.001 mm long). Narrow bands of chalcopyrite on the boundary between pink spar and sphalerite are likewise believed to be replacement of the pink spar, for the boundary with this mineral tends to be minutely irregular, whereas the opposite boundary, against the sphalerite, is sharp and even.

Nearly euhedral chalcopyrite crystals are imbedded in, and apparently have replaced, jasperoid in a few places, notably along a small fault in the Shorthorn mine.

Chalcopyrite also occurs in a minor amount in caves lined with large calcite crystals, without associated sphalerite, galena, or dolomite. Marcasite and pyrite are common associates. The chalcopyrite may antedate the other minerals, or it may be zonally embedded, along with the iron sulfide minerals, in the large calcite crystals.

In an unusual and minor occurrence noted in several mines, chalcopyrite has formed through the breakdown of earlier enargite. It builds up as a rough amorphous coating, starting usually from the bases of certain of the larger enargite crystal laths and extending in some cases to the tops of the crystals. The enargite is dissolved out and replaced although the process is not strictly metasomatic, for the final product may contain one or two hollow tubes in the center or at the sides of the crude pseudomorph. Furthermore, the base is commonly broader than the original crystal. In the replacement the two opposite edges and the base of the enargite lath are favored, and with complete removal of the host mineral, the final product is commonly a crude U-shaped skeleton of the original crystal, which appears to be built of many small botryoidal pellets of chalcopyrite (fig. 26). The two arms of the U are commonly hollow tubes. The surface may be black, presumably due to a thin film of chalcocite that has formed since the mine was opened.

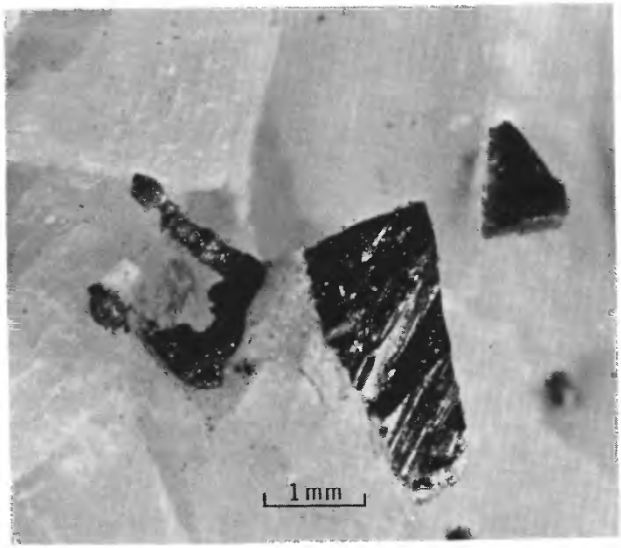


FIGURE 26.—U-shaped secondary chalcopyrite from Midcontinent mine. Originally formed by replacement along base and two edges of a lath-shaped enargite crystal which has leached out, leaving the chalcopyrite. Upper prong of the U stands as unsupported column; lower prong on dolomite has had a piece below the end broken out. Crystals on right are chalcopyrite.

Polished section study shows that the chalcopyrite, in its most common occurrence as a vug mineral, is extensively replaced by marcasite and subordinate pyrite, particularly near crystal surfaces. The product in extreme cases may approach a pseudomorph of the chalcopyrite crystal (fig. 27), but usually there is ample chalcopyrite left interstitially between the marcasite or

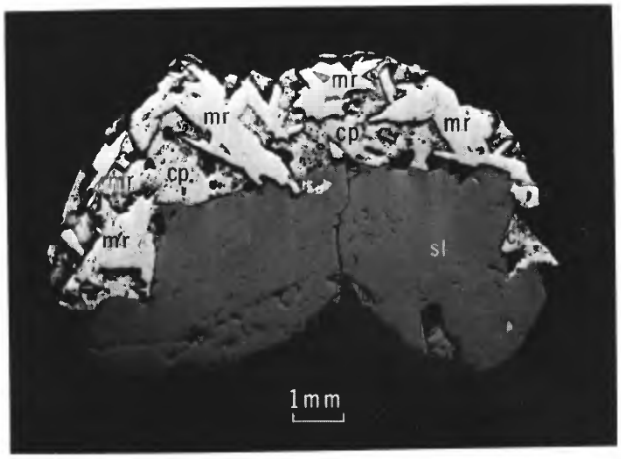


FIGURE 27.—Photomicrograph of polished surface showing extensive pseudomorphic replacement of chalcopyrite by marcasite from Midcontinent mine. Pyramidal external form of chalcopyrite shows in the peak at left. Chalcopyrite, cp; marcasite, mr; sphalerite, sl.

pyrite grains to show the nature of the original crystal that has determined the external form of the intergrowth. The replacing sulfide may have euhedral boundaries or an irregular but sharp anhedral boundary in the host mineral, or pyrite may have an indefinite wispy boundary, like frayed lace. Well-formed marcasite crystals that appear to grow on top of chalcopyrite nearly always have an irregular replacement root imbedded deeply in the underlying mineral.

That the replacement closely followed the deposition of the chalcopyrite is suggested in at least one specimen by the occurrence of small pyrite grains inside crystals of chalcopyrite that were formed on a "pause surface" about one-eighth of an inch in from the free surface of a galena cube. In general, however, the chalcopyrite enclosed in sphalerite or galena—and thus protected—is free of pyritic replacement.

Chalcopyrite has been the most susceptible of the sulfides to oxidation after the ore was exposed to air through mine openings. In many places the crystals are coated with brown iron oxide or basic sulfate, whereas closely associated sphalerite, marcasite, pyrite, enargite, and luzonite are unaltered or nearly so. Many such coated crystals, when broken, are found to be honeycombed with similar alteration products, and the surrounding rock may have sparse malachite stains.

Enargite (Cu_3AsS_4; Cu *48.3 percent*, As *19.1 percent*, S *32.6 percent*)

Enargite is found in the same geologic environment as chalcopyrite but is a less common mineral. Thus, whereas it always has chalcopyrite associated with it, the reverse is not true. Enargite is widely distributed in the field. If the fringe areas of the field had been more completely mapped, it is possible that enargite would prove to be nearly as widespread as the zinc mineralization. However, in the mines lying along and near the axis of the Miami trough between the Scammon Hill and Gordon No. 3 mines, nearly all of which were mapped, only two occurrences of enargite were found, one each in the Anna Beaver and John Beaver mines. Its absence can be correlated with the lack of dolomite and chalcopyrite in the southwestern part of this stretch but not in the northeastern part.

Within the known area of enargite occurrence, this mineral is much more abundant in the Kansas part of the Picher field, particularly in the Barr and Webber mines and in the group of mines extending from the Boska and Wilbur north through the Robinson and Tulsa Quapaw. Mines farther north in the latter group were not mapped, but as the Robinson and Tulsa

Quapaw contain much less than the Midcontinent, Bendelari, and Boska, possibly the main concentration of enargite comes in the neighborhood of these latter three mines. As a rough indication of its greater abundance in Kansas, 101 out of 136 enargite localities recorded in the mines mapped were in Kansas and the remaining 35 in Oklahoma.

Enargite forms tabular (fig. 28), lath-shaped, or, rarely, prismatic crystals that are usually bright and shiny. They may be scattered singly, or several crystals may be clustered on chalcopyrite, pink spar dolomite, or sphalerite; or rarely, the enargite crystals may be included in an outer zone in calcite crystals. Some are compound groups of 2 or 3 crystals in parallel orientation growing up from a common base. The largest crystals found are tabular, one from the Webber mine having dimensions of about 3.2 by 2 by 1 mm. An occasional crystal may be longer than this but not so broad or thick. Crystals growing on a chalcopyrite sphenoid, even though not in contact with each other, may have a parallel orientation controlled by the crystal lattice of the chalcopyrite.

A minor occurence of enargite shows in polished section as minute irregular blebs or bands growing in chalcopyrite which it probably replaces, usually on a grain border or around a pyritic inclusion. The boundary may be smooth or minutely serrate, with a certain amount of parallelism showing in segments of the tooth boundaries, suggesting that crystal directions in the chalcopyrite have controlled the tooth boundaries. In one unusual occurence (Midcontinent mine) the enargite forms an irregular veinlet included within a triangular crystal of chalcopyrite which, in turn, occurs inside a crystal of black jack sphalerite. One side of the veinlet is serrate and the other, smooth. A persistent direction of tooth edge in the serrated side is parallel to one of the outer crystal boundaries of the chalcopyrite. Other enargite that occurs just within the outer rim of the chalcopyrite crystal has the irregular or serrate boundary against the chalcopyrite and a straight outer boundary against the sphalerite, on line with the chalcopyrite crystal boundary.

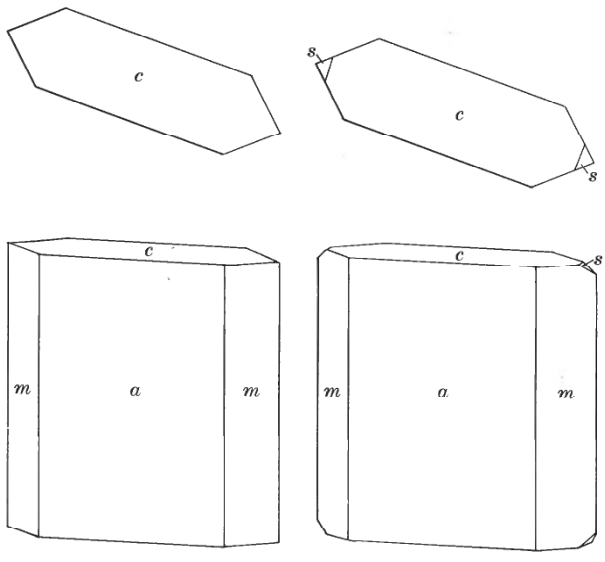


FIGURE 28.—Characteristic crystal forms of enargite in Picher field.

Most of the enargite blebs, when observed in polarized light, are found to be single crystals. A few, however, are made up of several interlocking grains or of two or three narrow parallel wedge-shaped grains in a larger host mass. There is no extensive fine lamellar twinning characteristic of luzonite (Ramdohr, 1950; Gaines, 1957). Furthermore, the color in plain light is gray (lighter than sphalerite) but with a tinge of lavender, which is more nearly the color of enargite.

In a few localities the replacement relations are reversed and their is undoubted partial replacement of isolated enargite crystals by chalcopyrite. In these crystals, however, the chalcopyrite is an unusual amorphous variety (p. 108). Only part of the enargite is replaced; the rest is either dissolved (leaving typically a U-shaped mass of chalcopyrite, fig. 26) or is recrystallized to an intermeshing network of fine shiny enargite crystals and luzonite skeletal crystals, commonly loosely suspended between the arms of the U. Generally the enargite replaced is that growing on pink spar. However, at one locality in the Grace Walker mine the replaced enargite crystals are perched on the earlier sphenoidal chalcopyrite. Some of this chalcopyrite, including that carrying replaced enargite, is strongly corroded to bright untarnished remnants, whereas closely adjacent crystals containing no enargite have not been attacked in the slightest degree (fig. 29). Although it cannot now be established that all the corroded chalcopyrite crystals were originally in contact with enargite, this would appear to be a possibility. The corrosion is possibly an electrochemical effect produced by the contact of the two minerals (Gottschalk and Buehler, 1912).

Spectrographic analyses of four enargite samples have been made by Janet D. Fletcher, of the Geological Survey. Each sample consisted of many small crystals carefully picked from a specimen from the indicated locality. Determinations of Sb and Ge were quantitative; determinations of other elements, only semiquantitative. The results are given in table 7.

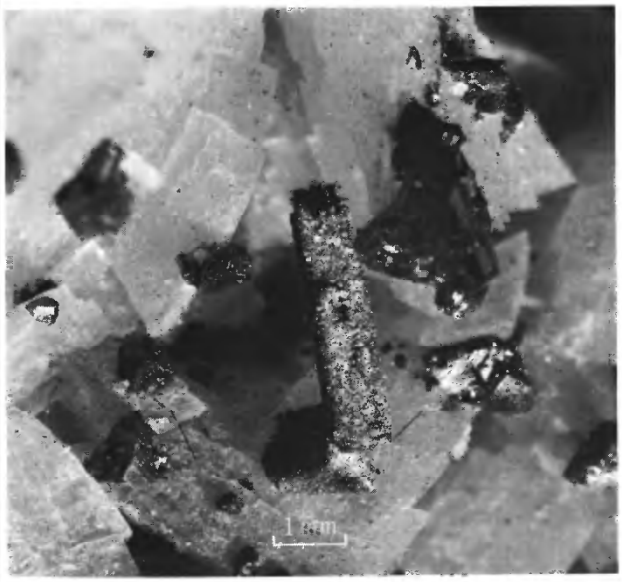


FIGURE 29.—Pseudomorph of chalcopyrite after enargite on corroded chalcopyrite crystal from Grace Walker mine. Adjacent chalcopyrite crystals have not been corroded. Specimen presented by Curtis Templain.

TABLE 7.—*Spectrographic analyses of enargite, in percent*

Sample	Ge	Sb	Ag	W	Sn	Zn	Ti
1	0.03	1	0.00x	0.0x	0.0x	0.x	0.00x
2	.04	2	.00x	.0	.0x	.x	.00x
3	.04	1	.00x	.0	.0x	.x	.00x
4	.04	1	.00x	.0x	.0x	.x	.00x

Looked for, but not found: Au, Hg, Rh, Pd, Ir, Pt, Bi, Mo, Cd, Tl, Co, Ni, Ga, In, Cr, V, Sc, Y, Yb, La, Zr, Th, Nb, Ta, Be, U, B. The samples were very small and were diluted to obtain accurate determinations for Ge and Sb. Owing to the dilution, some of the less sensitive elements may have been missed, and lead and iron could not be determined. Lead, if present, would be less than 0.1 percent.

1. Lawyers mine, NE¼NW¼ sec. 14 (a partial section), T. 35 S., R. 23 E., Cherokee County, Kans.; 430 ft south of north line, 510 ft east of west line.
2. Sample leached out of calcite crystal, Bendelari mine, SW¼SE¼ sec. 11, T. 35 S., R. 23 E., Cherokee County, Kans.; 75 ft north of south line, 370 ft east of west line.
3. Barr mine, SW¼SW¼ sec. 7, T. 35 S., R. 24 E., Cherokee County, Kans.; 510 ft east of west line, 360 ft north of south line.
4. Webber mine, NE¼SE¼ sec. 12, T. 35 S., R. 23 E., Cherokee County, Kans.; 60 ft west of east line, 1,100 ft south of north line.

Luzonite (Cu_3AsS_4; Cu *48.3 percent*, As *19.1 percent*, S *32.6 percent*)

This mineral, with virtually the same composition as the enargite, is quite distinct in its form, color, and crystal lattice. It has the same type of occurrence as enargite, and the two minerals are commonly found together, though each may occur independent of the other. The luzonite is less abundant bulkwise and has been found at fewer localities. These are chiefly on the Kansas side of the State line, in the same areas of concentration as outlined for enargite.

Luzonite from the Picher field has no crystal form, but appears as bladed crystallites perched on earlier formed minerals in vugs. The only crystallographic elements developed are directions of growth. A straight arm forms the base of the crystallite and is the thickest part of the grain. The point of attachment is at or near the middle of the arm, which obviously grew about equally in opposite directions from the point of origin. From the basal arm, the rest of the crystallite has grown as an irregular triangular but definitely pointed blade, the other two sides of the triangle beveling to relatively sharp edges. The blade is irregularly grooved by growth lines perpendicular to the base, and the thickest part is the growth segment containing the point of the blade. This segment may lie at the center of the basal arm or somewhat closer to one end; it is undoubtedly the earliest part of the blade, and grew from the point of attachment probably simultaneously with the basal arm, the rest of the blade then filling in the angles between these first-formed arms.

The overall shape of the crystallite, when viewed laterally, resembles that of a distant single-masted sailing vessel (fig. 30). This resemblance is further accentuated by the common placement of the blade point toward one end of the base (suggesting a mast that is fore of midship), by the usual concavity of the two "sail" edges, and by the irregularities of these edges. Uncommonly, a crystallite blade will have two points, suggesting a two-masted schooner, or the profile may be even more jagged.

In addition to the single crystallite blades, many specimens have a second blade starting from the same

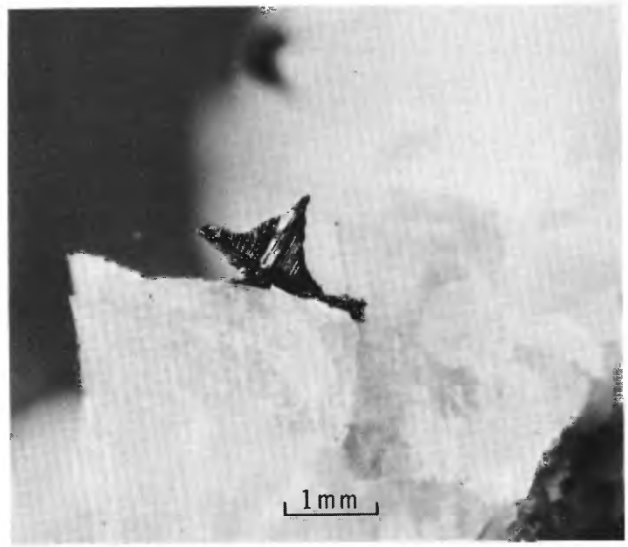


FIGURE 30.—Crystallite of luzonite perched on dolomite from Lawyers mine.

basal arm and extending at an obtuse angle that differs by perhaps 20° from a right angle; in some specimens there is a third blade from the same base and opposite to the second. In such compound growths there may also be partial blades or flanges extending at right angles to the basal arm and forming a web between the respective "ship masts." Where a crystallite has started growing on a sharp edge of some older mineral crystal so that its growth is uninhibited, some very complex bristly growths result.

These crystallite blades are 1–2 mm long and usually about half as high, though at some localities the height is as great as the length. They have a metallic luster, and are dark gray but have a distinct pinkish or bronzy tinge in comparison to the associated enargite.

Microchemical and qualitative spectrographic analyses indicate that the mineral is a copper-arsenic sulfide, similar to enargite in composition. A spectrographic analysis, quantitative for Sb and Ge and semiquantitative for several other elements, was made by Janet D. Fletcher, of the Geological Survey, on a sample made up of many small crystallite fragments carefully picked from the same specimen that yielded one of the enargite samples previously reported (table 7, No. 3). In table 8 the analysis is compared with that of the enargite from this same locality.

TABLE 8.—*Spectrographic analyses of luzonite and enargite, in percent*

Sample	Ge	Sb	Ag	Sn	Zn	Tl
A (luzonite)------	0. 10	1. 2	0. 00x	0. 0x	0. 0x	0. 00x
B (enargite------	. 04	1. 0	. 00x	. 0x	. x	. 00x

Looked for, but not found: Au, Hg, Rh, Pd, Ir, Pt, W, Bi, Mo, Cd, Tl, Co, Ni, Ga, In, Cr, V, Sc, Y, Yb, La, Zr, Th, Nb, Ta, U, Be, Sr, Ba, P, B. Owing to the dilution to obtain accurate determinations for Ge and Sb, some of the less sensitive elements may have been missed, and lead and iron could not be determined. Lead, if present, would be less than 0.1 percent.

A. Luzonite crystallites; Barr mine, SW¼SW¼ sec. 7, T. 35 S., R. 24 E., Cherokee County, Kans.; 510 ft east of west line, 360 ft north of south line.
B. Enargite crystals; same locality.

The small percentage of antimony in the luzonite is comparable to that in the enargite, and slightly lower than that found (1.48 percent) in luzonite from the type locality in the Philippines (Gaines, 1957). The germanium content of the luzonite is higher than that in the enargite, and comparable to that in some of the sphalerite from the Picher field (p. 103).

The X-ray picture made of the powdered mineral by Fred A. Hildebrand, of the Geological Survey (written commun., 1954), shows a pattern that is nearly identical with those of synthetic and natural luzonites, the latter from Luzon, Philippines, as described by Gaines (1957). This pattern is very different from that of enargite but closely similar to that of chalcopyrite (Gaines, 1957, p. 773).

Luzonite occurs most commonly on chalcopyrite but also on sphalerite and pink spar dolomite. At two closely adjacent localities in the Barr mine, luzonite perched on chalcopyrite crystals exhibits a definite orientation in relation to these crystals. The basal arm of the crystallite blade is closely attached parallel to a sphenoid face of the chalcopyrite crystal, and the blade extends out parallel to an axial plane of this crystal. Eight or ten blades in parallel orientation may appear on the same crystal face (fig. 31). As there are no more than one set of blades on a given face, it is probable that the axial plane involved is that of the horizontal axes when the chalcopyrite crystal is oriented in conventional position; for, in a tetragonal crystal, the two vertical axial planes are identical in symmetry and one would expect two sets of luzonite blades on a given face if the vertical planes controlled their orientation. Because the chalcopyrite crystals are heavily coated with oxidation products of iron, crystallographic measurements to determine this point can not be made.

Luzonite is most commonly clustered around the base of enargite crystals but rarely on such crystals, apparently because it is generally somewhat earlier than the enargite. Specimens were observed in which a delicate arm of the luzonite supports a much larger prismatic crystal of enargite on its distal end. Where an early phase of enargite has been destroyed and replaced by a secondary form of chalcopyrite, part of the original

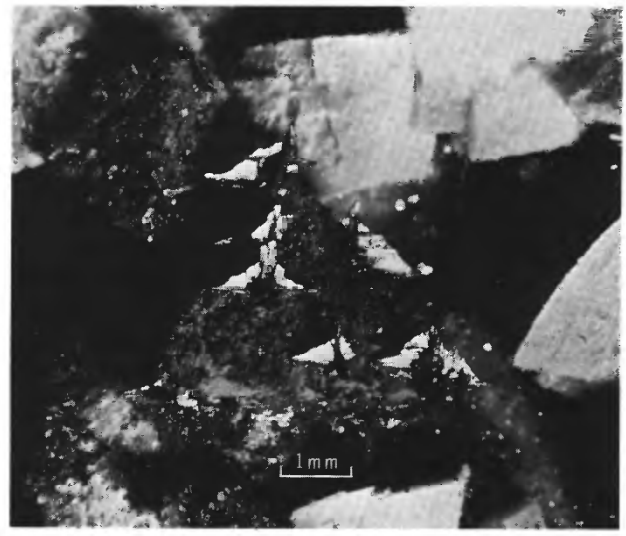


FIGURE 31.—Crystallites of luzonite reflecting light in parallel orientation on a limonite-coated chalcopyrite crystal whose lower sphenoid edge is horizontal in this picture. From Barr mine.

mineral has been recrystallized to an intermeshing network of enargite and luzonite (p. 110). Where this has happened, the luzonite crystallites show very little blade, for they are mostly the basal arm with an unusually long vertical arm ("mast") to which may be attached several cross spars parallel to the basal arm.

Marcasite (Mundic—FeS_2; Fe 46.6 percent, S 53.4 percent)

Marcasite is one of the commonest minerals of the district and occurs in a variety of forms. Most widely distributed are tabular prismatic and diamond-shaped crystals, usually a fraction of an inch long, but some have a maximum length of about 1 inch. These crystals differ from each other mainly in habit (prismatic laths elongated parallel to the *c* axis as contrasted with tabular blades flattened parallel to the side pinacoid), for the crystal faces are much the same. Fishtail twins are common. Repetition of fine diamond-shaped crystals in parallel growth produces a crude saw-blade type. Other types include a coarse cockscomb form (Palache and others, 1944, p. 312, illus.), which bulks large in some of the mineralized ground, and the following additional types which are quantitatively unimportant: narrow blades and needlelike forms of poor faces and undetermined orientation; small needlelike shafts of six-sided cross-section and pyramidal termination; a delicate hair form, resembling millerite, which may occur singly or in tufts; a rare massive type showing fine polysynthetic herringbone twinning, bordering massive pyrite of which it appears to be a replacement adjacent to vugs; and a microscopic crystalline but anhedral type replacing chalcopyrite.

The hair form is commonly an extension of needle and bladed forms along crystallographic axial lines. Such extensions are parallel to the faces that show striations and perpendicular to such striations. If it can be assumed that the striations are on the 010 face and in the zone 110 v 010, then the hairs are parallel to the *a* crystallographic axis. Some of the needlelike shafts and fine prisms, which may be only slightly coarser than the hairs, are obviously elongated parallel to the *c* axis.

Much of the marcasite, including the hair type, contains traces of nickel that can be demonstrated by microchemical methods. The best tests were obtained on small herringbone twin crystals associated with finely crystalline pyrite from one locality in the John Beaver mine. A small sample analyzed by R. C. Wells of the Geological Survey had a nickel content of 0.038 percent. The associated pyrite contained somewhat more nickel. Polished sections show that the marcasite overlies and encloses the small pyrite grains; hence, only the distal ends of the marcasite crystals were used in the sample analysed.

Barrett (1940) reports somewhat larger amounts of nickel, 0.06–0.09 percent, in specimens of marcasite from the Picher area. He also reports 0.03 percent copper and 0.3–0.5 ounce silver per ton, which is three to five times the amount of silver reported for the galena.

Marcasite is associated with the ore minerals throughout the field and is particularly conspicuous where it coats galena preferentially. In many pink spar vugs it shows a marked tendency to mass on the distal edges of the pink spar crystals. However, the major part of the marcasite is in jasperoid rather than dolomite ground; its greatest concentration is on the outer edges of ore bodies where the ore grades into barren rubble ground and boulder ground that had been leached of limestone prior to mineralization. A massive form coating the jasperoid and residual chert and bristling with coarse cockscomb crystals is particularly characteristic of such positions. Where the marcasite extends beyond the limits of the zinc-lead mineralization, it may occur alone, or be associated with, crystalline calcite. The large dogtooth calcite crystals found in caves beyond the ore bodies commonly have inclusions of marcasite concentrated in certain zones of the crystals. Another occurrence, probably unrelated to the mineralization, is as thin blades in the clay shale deposited in some of the underground caves.

Marcasite also occurs as microscopic disseminated crystals (diamond-shaped or prismatic) in jasperoid, but is much less abundant in this occurrence than pyrite, with which it may be associated. It may also be similarly disseminated in limestone that has been recrystallized and only partially replaced by jasperoid.

At one locality in the Kenoyer mine (NE¼SW¼ sec. 20), diamond-shaped marcasite crystals form a thick continuous coating on rounded crystals of ruby jack. Most of the ruby jack has leached out, leaving the marcasite as hollow shells. The smaller shells tend to be hemispheric, but the larger ones, as much as 4 or 5 inches across, are more flattened, resembling the broken-off tops of skulls. Such shells may show coalescence of several units of different sizes (fig. 32). Similar shells from which the underlying sphalerite has been partly or completely leached are found in the Barbara J and John Beaver mines.

At the Zig Zag mine, near Joplin, massive pyritic iron sulfide occurs interstitially between stalactites of the peculiar form of sphalerite (containing some wurtzite) that appears there. It may also be scattered irregularly inside the zinc sulfide masses, or, more rarely, interlaminated concentrically with such material. Polished

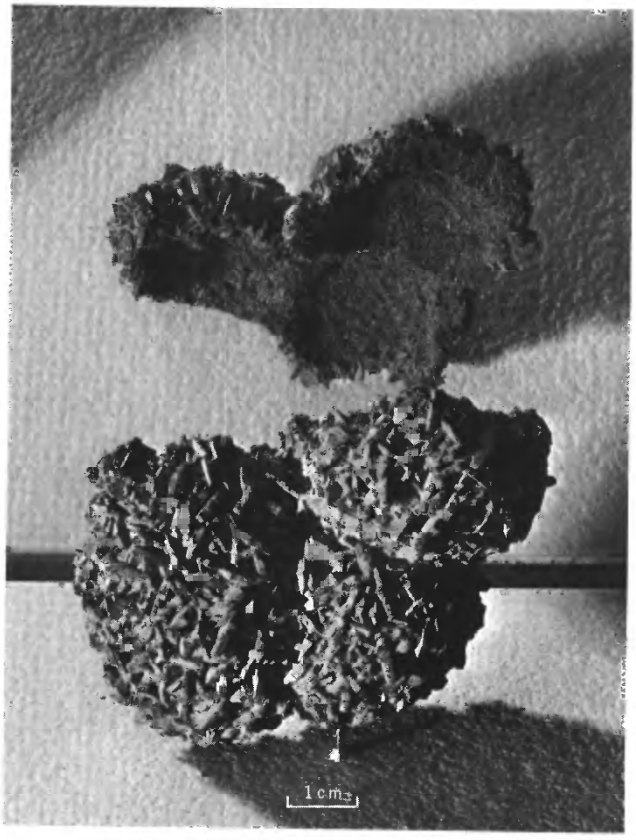


FIGURE 32—Shells of crystalline marcasite originally grown on sphalerite which has been leached from under the marcasite. From Kenoyer mine. Picture taken with a mirror to reflect underside.

sections show that most of this massive iron sulfide is pyrite, but it contains irregular angular grains of marcasite, which, in at least one specimen, are concentrated just outside a thin zinc sulfide shell that bristles with wurtzite crystals. Where the pyritic material borders vugs, it is commonly botryoidal, its free surfaces completely covered by small crystals of marcasite. Elsewhere, marcasite crystals form directly on the sphalerite surfaces. Such crystals are similar to the diamond-shaped and thin cockscomb forms found in the Picher field, but they contrast in that some are twinned on the less common marcasite law, with twinning plane on 011. In the more massive material bordering the vugs, there are numerous examples of polysynthetic twinning of the more common type, producing the herringbone pattern. Here again, there is a suggestion that such material may represent a replacement of massive pyrite.

Some of the friable mineralized jasperoid at the Zig Zag mine contains abundant microscopic marcasite (diamonds and flat prisms 0.02–0.10 mm long), in contrast to the usual dominance of microscopic pyrite in other jasperoids here, but more particularly at Picher.

Pyrite (*Mundic*—FeS_2; Fe *46.6 percent*, S *53.4 percent*)

Pyrite occurs usually in smaller crystals than marcasite or chalcopyrite. Although found throughout the Picher field, it is not abundant as a megascopic constituent of the ores. It is, however, an ubiquitous microscopic constituent of the jasperoid gangue.

Pyrite appears in the vugs and other openings as small cubes or pyritohedrons or as a combination of these two. The unmodified pyritohedron is much less common than the cube or the combination, and furthermore, it is generally associated closely, or even intermingled with, crystals showing one or both of the other crystal habits. The cube or the cube-pyritohedron combination, on the other hand, commonly occurs alone, only one crystal habit showing at a given locality. The maximum size of crystal is about a quarter of an inch in diameter, though most are pinhead size or smaller. There are enough examples in which the cubes are larger than the associated pyritohedrons to suggest that the cube may represent a later phase, grown from the other. Ramdohr (1950, p. 573) states that in zoned pyrite crystals, the crystal habit may change during the growth from octahedral in the interior to predominantly cubic in the outward parts.

As pyrite is late in the mineral sequence, it usually occurs on other minerals and shares with marcasite the peculiarity of commonly being concentrated on the distal points and edges of older minerals, particularly pink spar, but also sphalerite.

The pyrite disseminated in jasperoid gangue or in recrystallized limestone partly replaced by jasperiod has in part the same crystal forms as in the vugs, but, in addition, the octahedron is a common and characteristic form.[8] The octahedron may occur alone or may modify the cube, or, more rarely, the pyritohedron. All disseminated crystals are of microscopic size. Much of the finely disseminated material, however, is in grains of indefinite but usually rounded shape, which may show one or two crystal faces. One polished section (Kenoyer mine) contains spheroidal aggregates, 0.005–0.02 mm in diameter, made up of many discrete pyrite granules in typical framboidal texture (Rust 1935; Bastin 1950). Through varying degrees of coalescence in the granules, such aggregates show all gradations to rounded pyrite grains of comparable size, some of which may show one or more crystal faces (fig. 33).

[8] Crystal form of the pyrite was studied by mounting the crushed jasperoid in glycerine on a glass slide; in such a mount, the pyrite grains can be rolled by pushing the edge of the cover glass.

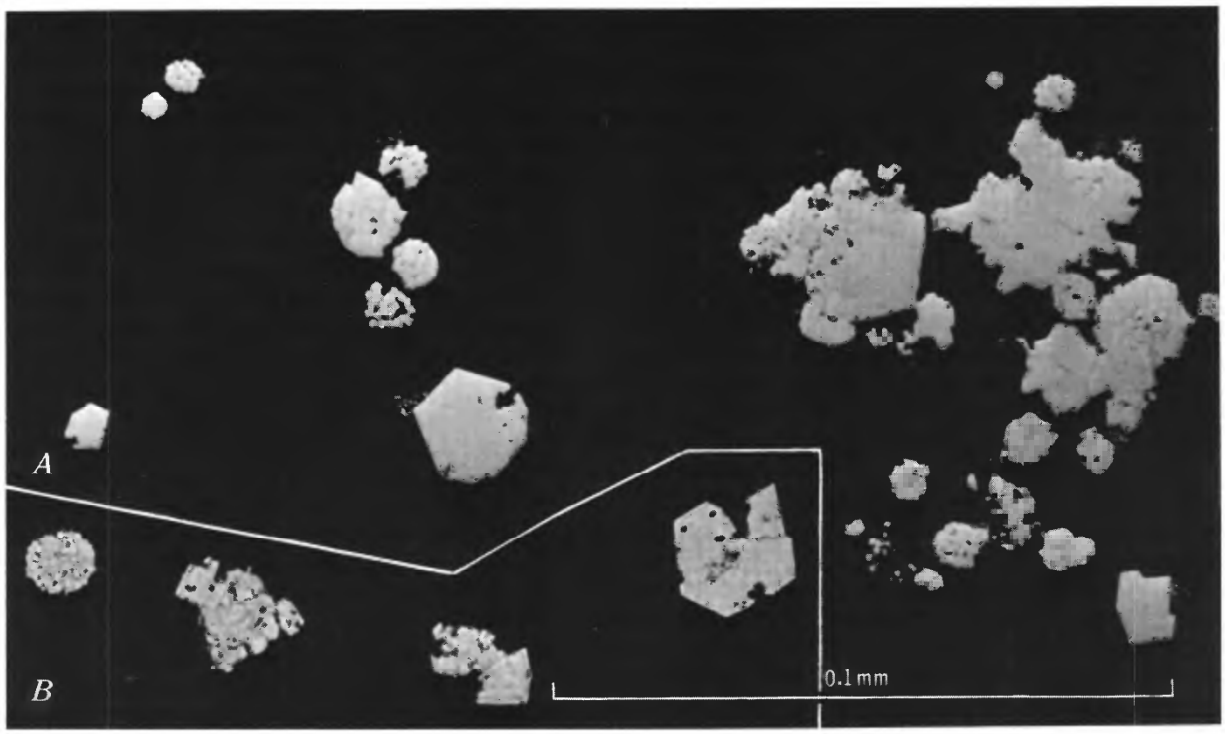


FIGURE 33.—Photomicrograph (two photographs are combined) of polished surface showing pyrite, with framboidal texture, disseminated in jasperoid. Pyrite coalesces in part to crystal outlines. From Kenoyer mine.

Disseminated pyrite grains and crystals of the type appearing in jasperoid are also found but are less abundant in cotton rock and chert. Their occurrence in massive gray spar that replaced limestone early in the mineralization sequence is erratic, and may be correlated with the jasperoid that is usually present. Pyrite similar to that in the jasperoid is also found in the fine-grained sandstones and shales of Pennsylvanian age that fill caves in the ore deposits; this pyrite may well have been introduced at the time of the lead-zinc mineralization.

Pyrite also commonly occurs as massive replacement blebs in jasperoid, gray spar dolomite, and chert, and locally, such blebs may be several inches across. Along the Velie Lion-Anna Beaver boundary line, massive pyrite several feet across occurs along the fault zone on the south side of a circular slump pipe. This massive material contains striated cubic crystal faces as much as 1 inch across. On the borders the pyrite grades into jasperoid of the surrounding ore-bearing block.

Pyritic replacement of chalcopyrite may be in small cubic crystals, but is more commonly anhedral, some of the masses being extremely ragged in minute detail because of the incomplete replacement that leaves inclusions of remnant chalcopyrite intergrown with the pyrite (fig. 34).

Some of the pyrite contains enough nickel to be indicated by microchemical tests. Best tests were obtained on minute cubes (0.3–0.5 mm in diameter) scattered over a chert surface at a locality in the John Beaver mine. A small sample of this material, analyzed by R. C. Wells, of the Geological Survey, contained 0.18 percent nickel. A polished section shows barely perceptible zoning in the crystals in that the centers have a slightly grayer tinge in comparison to the rims. The centers also show very slight, but definite, anisotropism which is distinguishable chiefly because of its hourglass distribution (fig. 35). Neither the zoning nor the anisotropism was noted in other pyrite, and it is probable that these features are due to nickel concentrated in the cores of these crystals which would thus carry a higher percentage proportionally than the overall content determined by the analysis cited above.

At the Zig Zag mine near Joplin, massive pyrite fills part of the interstices between the stalactites of zinc sulfide that are peculiar to this locality. It is also scattered irregularly within the zinc sulfide masses, or, more rarely, is interlaminated with such material. Some interlaminated pyrite is concentrically banded (fig. 36). A few small crystals of pyrite (cube + pyritohedron) are found on the stalactite surfaces of the zinc sulfide,

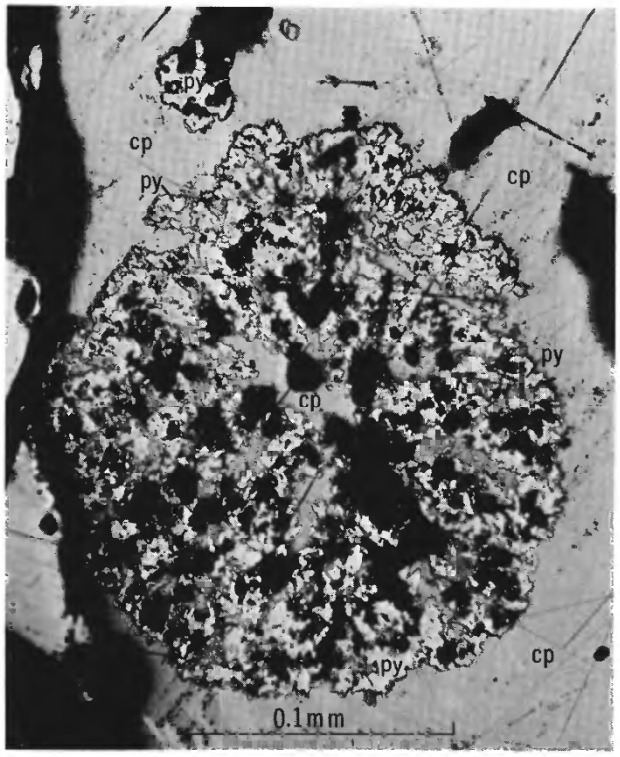


FIGURE 34.—Photomicrograph of polished surface showing replacement of chalcopyrite by pyrite (with ragged boundary) from Midcontinent mine. Black spots are pocks in the polished surface. Chalcopyrite, cp; pyrite, py.

though most of the iron sulfide in this position is crystaline marcasite.

Dolomite (*gray spar*, *pink spar*—$CaMg(CO_3)_2$)

Dolomite occurs in two forms, a gray granular replacement of limestone and a light-pink crystalline form (rhombohedron) that characteristically lines vugs but also fills fractures and irregular seams and pockets in the mineralized ground.

In the massive replacement by gray spar, the impurities in the limestone, mainly clay minerals and, locally, glauconite, are segregated interstitially between the gray spar grains. Conceivably, the calcium entering into each grain of the replacing dolomite came from the limestone that previously occupied the same spot. The grain size averages perhaps 1–2 mm across.

The pink spar was obviously precipitated from solutions moving in free openings. The contained calcium may have moved a considerable distance from its former location in the sedimentary limestone. Where pink spar occurs in replacement pockets in the gray spar or jasperoid, all interstitial impurities in the original limestone have been removed so that the pink spar pocket is, in effect, filling a void. The pink spar is usually closely associated with the gray spar, but locally, the solutions carrying the constituents of dolomite may have moved into fractures or vugs slightly outside the limits of the gray spar replacement. Although pink spar crystals vary in size, the size at any given locality tends to be uniform. Most commonly they are between 0.25 and 0.4 inch. The crystals have the usual pronounced curvature of dolomite to crude saddle shapes. When a given crystal has reached a size of about 0.4 inch across, it has usually become so curved that further growth separates as a new unit, slightly rotated from the first unit. In a few examples, however, a crystal may reach a maximum size of about 0.6 inch long without any conspicuous break in continuity, though such crystals invariably show, in the curved crystal faces, small offsets along rhombohedral surfaces.

The pink color is presumably due to a trace of manganese. Similar pink spar from the Red Cloud mine in northern Arkansas contains 0.08 percent manganese and 0.90 percent iron (McKnight, 1935, p. 111).

Where dolomite of either type is replaced by jasperoid, the remnant grains tend to retain the rhombic cleavage form, even though they may be partly embayed by grains of the replacing jasperoid.

Although dolomite is widespread in the Picher field, its distribution is uneven and it is absent from large

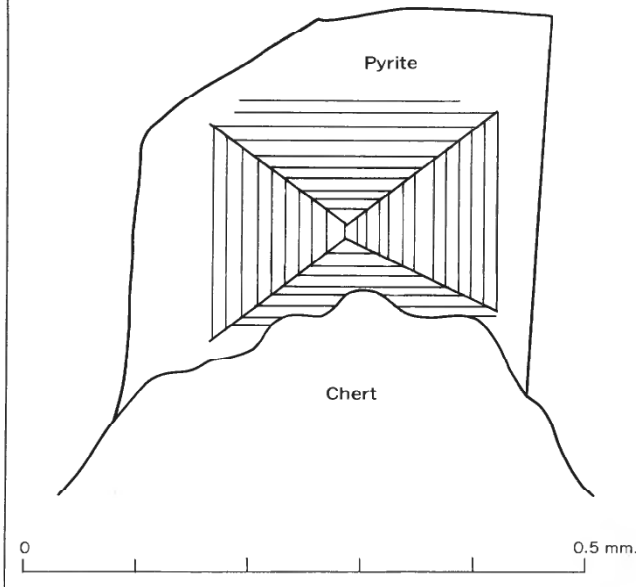


FIGURE 35.—Sketch showing anisotropic mass in center of nickeliferous pyrite crystal from John Beaver mine. The horizontal hourglass figure is an optical unit contrasting with another optical unit comprising the other two quadrants.

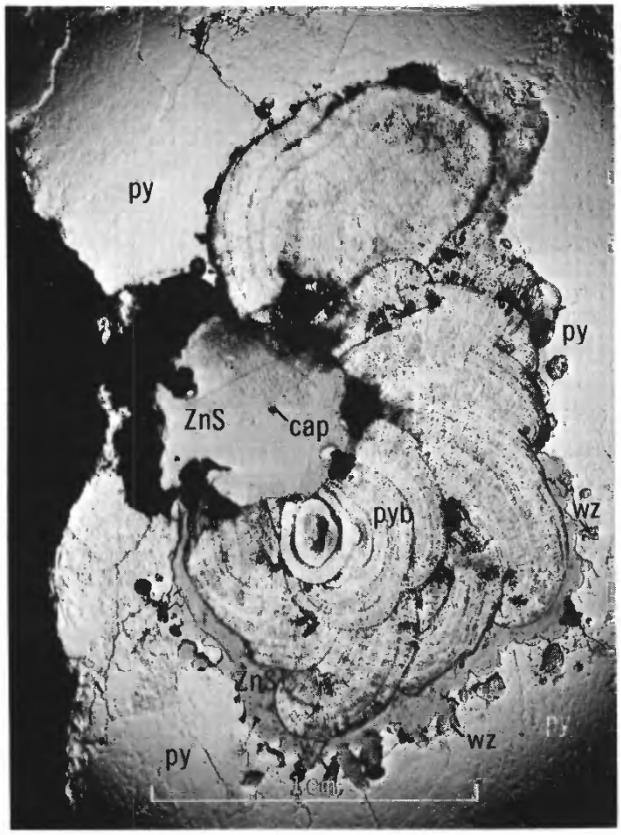


FIGURE 36.—Photomicrograph of polished surface showing massive pyrite interlaminated with zinc sulfide in stalactite from Zig Zag mine, near Joplin. Note concentric banding in the pyrite, and capillary opening in the zinc sulfide. Massive pyrite (containing a few marcasite grains adjacent to the wurtzite zone), py; banded pyrite, pyb; massive zinc sulfide, ZnS; wurtzite crystals, wz; capillary opening, cap.

areas, particularly in the southwestern part of the field (pl. 9). Wherever dolomite does occur, it is usually the dominating gangue mineral in blocks of ground that may be as much as several hundred feet across.

Calcite (*tiff*—$CaCO_3$)

Calcite is a very common and widely distributed gangue mineral, occurring both as well-formed crystals in vugs and caves and as crystalline aggregate filling fractures, bedding seams, breccias, and replacement pockets in the mineralized and adjacent ground. Replacement pockets several feet long and 2 or 3 feet thick in the Velie Lion mine have cleavage faces as much as 2½ feet across. Calcite may also occur as an interstitial matrix in the more porous jasperoids, having crystal continuity in unit sizes of half an inch or more as shown by the reflection of light from cleavage surfaces. The amount of calcite is very irregular, from abundant in many workings to sparse or even absent in some stopes.

Most of the calcite is transparent and colorless, but many of the crystals are shades of yellow, grading to amber. The largest crystals may have a brownish-lavender or pinkish color in the center and a sharp transition to paler or colorless in an outer shell.

Most of the calcite crystals fall into three general habits: (1) prismatic with rhombohedral termination, (2) scalenohedral ("dogtooth"), and (3) flat rhombohedral. Although there is some gradation between these types, on the whole they are fairly distinct and separate. All three occur with the zinc and lead ores; the prismatic and scalenohedral types are about equally abundant and widespread in this association, whereas the rhombohedral form is less abundant.

The prismatic crystals show the prism m $(10\bar{1}0)$, terminated by the negative rhombohedron $\delta\cdot(01\bar{1}2)$ or $\eta\cdot(04\bar{4}5)$, or commonly by both. An additional modifying rhombohedron, $\phi\cdot(02\bar{2}1)$, is not common. In a rare variant found at one place in the Chubb mine and also in the Wilson mine, the prism is replaced by a steep negative rhombohedron $\Sigma\cdot(0\ 11\ \bar{1}\bar{1}\ 1)$. Prismatic crystals 1–2 inches long are characteristic of the ores, though smaller crystals, as small as a fraction of an inch long, are also plentiful. The largest prisms are about 6 inches long and 3 inches in diameter, and, in general, have poorer crystal faces than the smaller sized crystals.

The scalenohedral crystals show the usual form K: $(21\bar{3}1)$, commonly terminated by small faces of several other forms, including the unit rhombohedron $p\cdot(10\bar{1}1)$ and the same two negative rhombohedrons that terminate the prismatic habit. Very common modifying forms are the steep rhombohedrons $\phi\cdot(02\bar{2}1)$ and $m\cdot(40\bar{4}1)$, which, in the smaller crystals, may develop to the subordination of the scalenohedron, though the resulting overall shape is still "dogtooth." Common to these more complex crystals are inconspicuous line faces of three modifying flat scalenohedrons, namely F: $(41\bar{5}3)$, $71\bar{8}6$, and $71\bar{8}9$,[9] which lie adjacent to the unit rhombohedron. In certain rare complex crystals (Kansouri and Barr mines) the scalenohedron P: $(32\bar{5}1)$ takes the place of the more common form K: $(21\bar{3}1)$. At the Wade mine at the eastern end of the field, the calcite crystals depart from the usual "dog-

[9] I am indebted to W. T. Schaller, of the Geological Survey, for determination of these forms on a selected crystal. Recognition of the same and other forms in other crystals has been based on inspection and rough measurements of angles in a crude reflection goniometer, improvised by cementing the crystals in appropriate orientation on the hinge of a Penfield contact goniometer and reflecting a beam of light into the eye from crystal faces whose angles are to be measured.

tooth" habit in that the scalenohedron *K:* $(21\bar{3}1)$ is terminated by extensive faces of a flatter scalenohedron *w:* $(31\bar{4}5)$, producing a more rounded crystal. This habit is shown by 1-inch crystals associated with the ore as well as by smaller crystals capping crinoid stems that project from the walls of solution caves in limestone. The capping crystals in the latter instance are oriented with the *c* axis on the prolongation of the central axis of the crinoid stem. Crystals similar to those at the Wade occur at one locality in the Tulsa Quapaw mine, except that the flat scalenohedron is *t:* $(21\bar{3}4)$ instead of *w:* $(31\bar{4}5)$; these crystals are conspicuously twinned on the basal pinacoid.

Scalenohedral crystals range in length from a fraction of an inch to as much as 4 feet. Most of those found with ore are less than 6 inches long. At the other extreme are the huge crystals that occur in caves opened in mining. Such caves commonly contain neither zinc nor lead minerals, or so little as to be unprofitable for mining. Calcite is one of the last minerals to crystalize and does not overlap the period of crystalization of either sphalerite or galena. Hence, there is no deposition of these ore minerals during the prolonged growth of the calcite in such caves, and the only ore would be that present on the walls before the beginning of calcite deposition. As caves are common at and beyond the borders of the ore-bearing ground, these are the ones that evidently fill up with the coarse dogtooth calcite at some time after the zinc-lead deposition, if any, at such localities. It is not evident why the smaller caves and vugs in the ore are not filled by this coarse calcite at the same time.

The pink to purplish color appearing in the cores of many of the large crystals is virtually limited to these giant scalenohedral crystals. An exception was found at the west end of the Federal Jarrett mine, where prismatic crystals, of the maximum size recorded for that type, also have purple cores. As the large crystals must have grown from small ones, the nonoccurrence of such small purple-tinged crystals is puzzling. The pale or colorless peripheral zone in the giant scalenohedral crystals may be 1–2 inches thick.

The flat rhombohedral crystals have only the one essential form $\delta\cdot(01\bar{1}2)$. However, the equatorial edges of the crystals are commonly truncated by small faces of the prism. Well-formed crystals are rare and less than an inch across. Several crystals may be aggregated in random orientation to produce a rosette, or stacked with their vertical axes in line to produce a barrel- or corncob-shaped mass in which only the rhombohedral terminations are good crystal faces. Most of the "corncobs" are 3–6 inches long; the maximum is about 1 foot long. Their rough sides, formed by the coalesced acute edges of the rhomb, are opaque, whereas their cores, and usually the rhombohedral terminations, are clear. Botryoidal and kidney-shaped masses as much as 1 foot across, found in certain mines (Dobson, Jaybird, and Crawfish), are evidently an extreme development of the rosette form in which the individual crystals are small and multitudinous. The peripheral shells of such masses are generally opaque white to cream colored and may be concentrically banded, but the centers are clear. That the various forms of calcite herein described under the rhombohedral type constitute a homogeneous unit is indicated both by gradations in shape and by a close association and alternation of the various forms in space, showing that they developed under virtually uniform conditions.

There are other crystal habits that do not fit into the three common ones described, but such are not common. Many are modifications of the scalenohedral habit, in which the other faces mentioned under the description of that type have attained such prominence as to suppress the usual scalenohedral appearance. Another habit, found at adjacent localities in the New Chicago No. 1 mine, is pseudo-octahedral, from a combination of the flat negative rhombohedron $\delta\cdot(01\bar{1}2)$ with a steeper rhombohedron, probably $\psi\cdot(05\bar{5}2)$. Many other forms occurring as modifying faces throughout the mining field have not been accurately determined, but these do not affect the overall habit.

At a given locality, one type generally occurs to the exclusion of the others. The three main types alternate irregularly with each other throughout the studied part of the field, each occupying an irregular area of the underground workings that is usually at least several acres in extent, and in some places the long dimension of such an area may exceed half a mile. There may, however, be small islands of one of the other types in such large areas. Where two mine levels are superposed, the crystal habit of the calcite on the two levels may be different, even though the vertical separation may be only 50 feet. However, no extensive overlap of different types on different levels was observed.

The conditions that determine the crystal habit are not evident, although the size of crystal may be one small factor. At localities where the prismatic crystals prevail, the few crystals that may have modifying faces of the scalenohedron are likely to be the larger crystals; whereas in a scalenohedron environment, the crystals that may have modification by the prism, or by various rhombohedrons and flat scalenohedrons tending toward the prismatic shape, are likely to be the smaller crystals. It will be recalled, also, that the scalenohedrons attain

a much larger maximum size. These facts would suggest that, inasmuch as large crystals must grow from small ones, the scalenohedrons are later phases of the prismatic crystals. Such, however, is at variance with the few observations that have been made as to the relative ages of these two types.

In only a few places where different types occur at a given spot can age relations be worked out. On the Gordon No. 3-Anna Beaver line, nearly 1,000 feet from its east end, prismatic and scalenohedral crystals occur in close proximity, and at one place they are combined in the same crystals, so that scalenohedral points project from the ends of prismatic crystals. Ordinarily, it would not be possible to determine which is the later, but in these crystals the reflection of light from the faces $\phi\cdot(02\bar{2}1)$ beveling the edges of the scalenohedron can be seen through the transparent prismatic overgrowth, indicating that the prism is later than the scalenohedron and its modifying faces. Furthermore, the prismatic overgrowth is not even on all sides, but shows offsets in the terminal rhombohedral faces around the scalenohedral corners. Finally, there are tiny excrescences of the prism-rhomb phase on the scalenohedral faces above the level of the enclosing prisms. Similar excrescences, though on a much coarser scale, were observed on 4-foot scalenohedrons lining caves in the Anna Beaver mine (NE¼NE¼ sec. 19) and John Beaver mine (NW¼ SE¼ sec. 19), and again show that the prismatic type was later. In view of frequently observed reversals in order of crystallization between differing mineral species, it is perhaps unsafe to assume that the order between different types of the one species, calcite, is always the same.

The flat rhombohedral type (including the "corncob" and botryoidal modifications) is later than the scalenohedron in the few places where the relations are revealed, and is probably also later than the prismatic type, but it seems to be unstable in relation to these types. The general opacity of much of the material suggests granulation. In certain examples of "corncob" calcite from the Chubb mine, extensive solution in concentric zones just below the surface in the opaque layer has left an incomplete external shell over a clear scalenohedral core. The transparent core was in existence before the leaching, though its crystal form may have been perfected during the leaching. In the Jay Bird mine, opaque botryoidal masses of calcite, the surfaces of which are a "felt" of crystal (probably rhombohedral) edges, overlie scalenohedral crystals and show more or less leaching along the contact with these crystals, whereas the latter are unleached.

At numerous localities in the Gordon No. 3, Anna Beaver, John Beaver, and Velie Lion mines, the free ends of the prismatic crystals have axial extensions which are small replicas of the main crystals and in parallel orientation with them, resembling a cupola on a silo. This feature is common over an area that is somewhat more than a mile across. A similar feature, differing in that the "cupola" is a scalenohedral point extension, has been described from one locality on the Gordon No. 3-Anna Beaver line and has been found also in the Federal Jarrett and Barr mines. Possibly these latter occurrences likewise record an overgrowth of crystalline material of differing habit over an earlier scalenohedral core.

Jasperoid (SiO_2)

The microcrystalline form of quartz known as jasperoid is the most abundant and widespread of gangue materials. It occurs predominantly as a massive replacement of limestone and gray spar dolomite, but it also occurs in the dolomite, and less commonly, in the limestone, as replacement bands with gradational or sharp boundaries, both along and across the bedding and as indefinite wisps. Almost all mineralized ground contains some easily recognizable jasperoid, and incipient silicification by scattered jasperoid grains that can only be found by microscopic examination is ubiquitous. Limestone is much more susceptible to replacement than dolomite. To some extent the jasperoid also permeates and partially replaces cotton rock, chert, sandstone, and shale. In noncalcareous materials, however, it is not ordinarily possible to distinguish between direct replacement and the replacement of limy lenses that were originally present in such material. In large areas of mineralized ground, for example in the southwestern part of the field between the Blue Goose No. 1 and Southside mines, very little initial replacement by dolomite has taken place, and in such ground the quantity of jasperoid replacing all the original limestone between the chert bands, nodules, and breccia fragments is particularly impressive.

The jasperoid may be porous, friable, and earthy, resembling a clay, or more commonly, dense, hard, and translucent. It usually ranges in color from gray and brown to nearly black, but extremely porous material may resemble pale sand. Jasperoid that replaces the fine-grained brown limestone in the basal part of the Moccasin Bend Member of the Boone Formation (G and H beds) is commonly dense and of a chocolate-brown color.

The quartz grains composing the jasperoid, as seen in thin section, usually form a tight-fitting mosaic of anhedral grains, and the tendency for a large percent-

age of such grains to have prismatic elongation (fig. 37) is characteristic. The perfection of this elongation varies from place to place, but where it is most conspicuous, the grains are three to four times longer than broad. The prismatic grains are commonly 0.01–0.1 mm long, and may reach a maximum of 0.5 mm. The size of grain is quite variable, even in the same thin section. Where replacement of carbonate material is only partial, the scattered jasperoid grains develop euhedral boundaries against and inside the carbonate but retain the same approximate size as the larger of the more crowded grains. Later solution of the carbonate yields a very porous friable jasperoid, which can be so loosely felted together that its composing grains have slumped and become crudely stratified on the indefinite floors of porosity-slump caves. Euhedral grains of jasperoid may also occur in the edges of sphalerite and galena blebs that are disseminated in the jasperoid.

As pointed out in the discussion of marcasite and pyrite, scattered microscopic crystals of these minerals are typical of the jasperoid. Glauconite is retained unchanged from its occurrence as scattered grains in the original limestone, as are probably also various clay minerals and sedimentary mica, which occur both in minute scattered flakes and as aggregates of fine flakes in parallel orientation along certain shaly zones. Also fairly common at horizons above that of the Short Creek Oolite Member are minute detrital grains of tourmaline. Where jasperoid has replaced dolomite or limestone containing scattered crystals of dolomite, small remnant grains of the latter may be present and tend to show rhombic outlines in contrast to the anhedral shapes of residual calcite. The larger grains of jasperoid may occlude minute rounded remnants of calcite, or questionably dolomite, from the replaced rock.

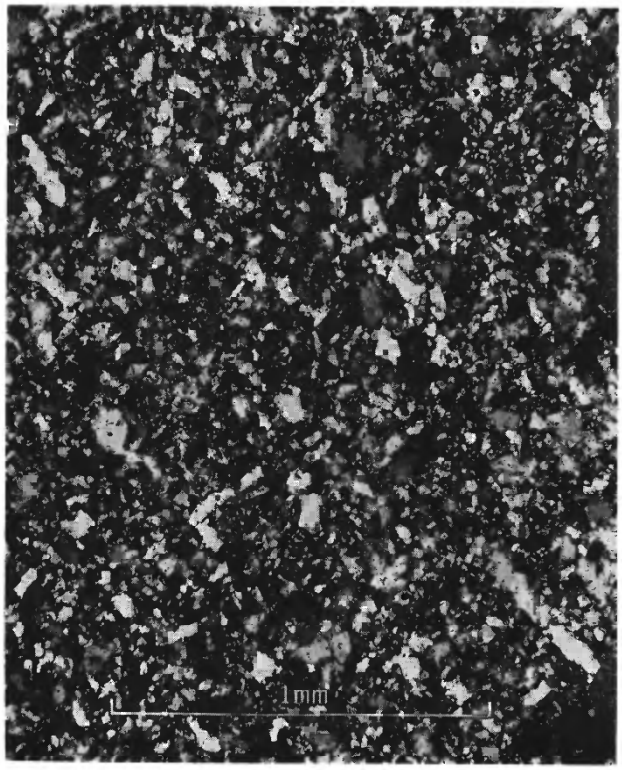


FIGURE 37.—Photomicrograph of thin section showing prismatic elongation of jasperoid grains in specimen from Robinson mine. Crossed nicols.

In replacement of the Short Creek Oolite Member by jasperoid at the Iron Mountain mine, many of the oolitic pellets are leached out, and the resultant holes are either unfilled, leaving a pumicelike product, or are filled by fine disseminated sphalerite.

Where jasperoid has partially replaced earlier siliceous rocks, such as chert, cotton rock, and sandstone, it appears as scattered microscopic grains in a matrix of the earlier material. In cherts and cotton rocks, which are composed of minute irregular quartz grains without consistent elongation, the jasperoid grains appear much larger and commonly have at least a little prismatic elongation, though nowhere near the perfection shown where the replaced material is carbonate rock. In sandstones the jasperoid is interstitial and not sharply separable from very small detrital quartz fragments.

Jasperoid that replaces comparatively undisturbed limestone in M bed is commonly banded parallel to the bedding, on a surprisingly fine scale when one considers the massive appearance of the unaltered limestone. At one locality in the Wilson mine, 12–15 obscure bands to the inch can be recognized in certain zones, and better defined bands, accentuated by fine disseminated sphalerite, occur at irregular coarser intervals averaging about an inch apart (fig. 38). At places where jasperoid is intimately interbanded with gray spar dolomite, as in parts of the Pelican mine, the pattern is probably inherited from an earlier banding of limestone by the gray spar; the jasperoid later replaced the limestone differentially. Such a sequence can be demonstrated in the North Arkansas zinc mines, where the obvious lamination of the limestone country rock has greatly favored the interbanding of such metasomatic products as dolomite and jasperoid (McKnight, 1935, p. 112, 119).

The fossils that were present in the original limestone, particularly the crinoid stems, are well preserved as casts in the more porous jasperiod. Calcite granules included in jasperoid grains are commonly arranged in arcs and wreaths that are undoubtedly relicts of microfossils.

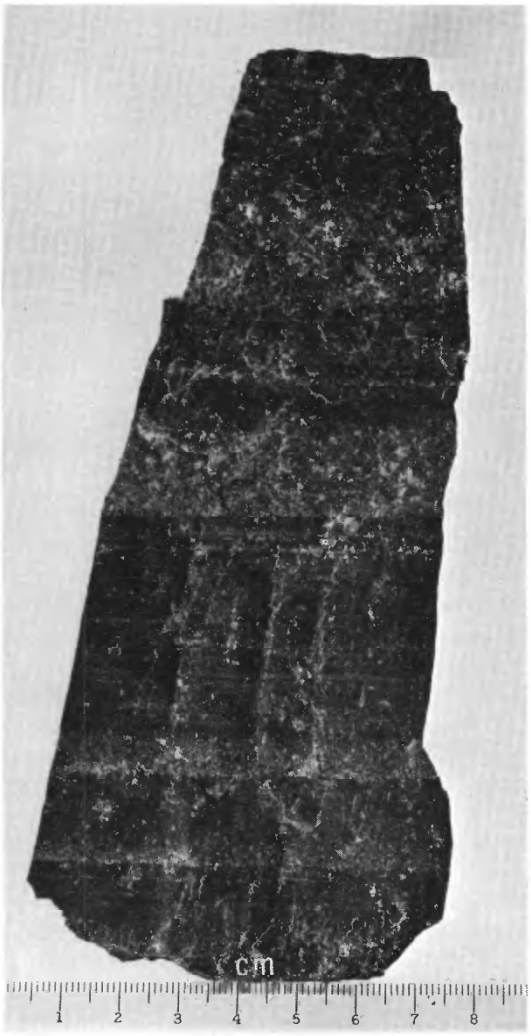


FIGURE 38.—Specimen from Wilson mine showing fine banding in jasperoid.

Although most of the jasperoid is obviously a replacement of limestone or dolomite, there are a few local features that suggest small-scale direct filling of open vugs and fractures. Nevertheless, most of these can best be explained as a replacement of some preexisting substance. The most puzzling examples show vugs that are partly filled by jasperoid up to a certain level, and are open above. Ignoring small-scale details, the upper surface of the jasperoid is remarkably smooth, and either horizontal or only slight undulant. In widely separated localities, such surfaces show an unusual pattern of parallel and semiparallel ridges and troughs of slight amplitude, resembling minute ripple marks in which the distance from crest to crest averages about 1 mm (fig. 39). As seen in cross section, the upper part of the jasperoid is denser than normal, and it may present a sharp lower contact with more typical jasperoid below. Microscopic examination shows that, except for being finer grained, this dense phase is very similar to the typical jasperoid. It shows the usual prismatic elongation of the quartz grains; it also shows the presence of minute flakes of sedimentary mica, an occasional glauconite grain, and small rhombs of carbonate, all of which are presumably residual from earlier limestone or dolomite that was replaced by the jasperoid. It is believed that such replaced material was originally a bedding band, but possibly the replacing jasperoid went through a colloidal stage, which resulted in some shrinkage under the influence of gravity before final solidification, leaving an open space at the top. Thin planar vugs are especially common at the tops of jasperoid bands and just below the interbedded chert bands in G and H beds.

Where jasperoid fills crosscutting fractures in chert, it has, for the most part, replaced limestone that had been plastically squeezed into the fractures at an earlier stage. Narrower veinlets of jasperoid may represent

FIGURE 39.—Specimen showing "ripple marks" on jasperoid surface on floor of vug, with pink spar crystals projecting above the surface. From Pelican mine. Upper figure is mirror image of lower, to show cross section.

replacement of the chert or cotton rock along a crack. Veinlets of jasperoid in chert, gray spar dolomite, or in jasperoid of a different color and texture commonly contain glauconite and sedimentary mica, which are evidence of replacement of an earlier sedimentary material.

Nevertheless, a few sharp-walled jasperoid veinlets a fraction of an inch thick in gray spar and in chert and a few matrices of fine chert breccias may represent a filling of open spaces. This jasperoid is dense and commonly dark. Another minor occurrence that apparently has not replaced an earlier sediment is a paper-thin film that has been deposited sparingly on the surfaces of galena crystals and, to a less extent, on sphalerite crystals in open vugs. It would not normally be noticed, except for an intense leaching of the galena at a later stage, which leaves the film standing as a conspicuous wall. Although the film is usually whiter than typical jasperoid, it locally thickens to a light-gray dense coating a small fraction of an inch thick, which resembles a light-colored jasperoid not only in its general appearance but also in the size and prismatic habit of the composing grains.

Quartz (SiO_2)

Quartz, as here distinguished from jasperoid, is a more coarsely crystalline form of silica which has crystallized in open vugs or cracks. The crystals of the Picher field are terminated by positive and negative rhombohedrons in nearly equal development, which thus simulate hexagonal pyramids. These pyramids may be the only crystal faces that are present on a crystalline surface, but commonly, there are also short prisms below the pyramidal points. Most of the crystals are small, the maximum diameter being about half an inch. The coarser quartz is found with ore in the Grand Falls Chert, where it may coat the chert walls of vugs or may underlie or coat the sphalerite and galena crystals. Individual crystal are clear, but in groups of small crystals they may appear white. Many fine quartz surfaces are no more than the free crystalline surfaces of jasperoid masses, and the distinction between the two phases of silica in such examples may be artificial.

Apatite [$Ca_5(PO_4)_3F$]

Some of the jasperoids, when examined in thin section under the microscope, contain minute colorless prisms or discs that have hexagonal cross section, high index of refraction, low birefringence, and negative optical elongation. Many are so small that they appear to be isotropic. There are gradations to anhedral grains which may or may not have crystal faces on certain sides. Some of this fine-grained more or less crystalline material is associated with rounded blebs of amorphous phosphate rock (collophane) known to be present as a sedimentary product in the unaltered limestones. The finer anhedral grains of the crystalline material are indistinguishable, because of apparent isotropism, from the smaller grains of collophane, and this, together with the typical crystal forms and optical properties cited, indicates that the crystalline material is apatite, which has the same composition as collophane. Whether the fine apatite crystals are present in the unaltered limestone has not been determined, but it would seem reasonable to assume that they represented a recrystallization of disseminated sedimentary phosphate brought about by the hot mineralizing solutions during the replacement of the limestone by jasperoid. On the other hand, the occurrence of apatite in mesothermal veins of the Alpine type is well known; hence, the phosphate could have been carried as a primary constituent of the ore solutions.

Kaolinite ($2H_2O \cdot Al_2O_3 \cdot 2SiO_2$)

A white pulverulent powder which occurs in small solution pores in the jasperoid matrix of a "headcheese" breccia at the Midas mine is composed of minute hexagonal and rhombic plates that can be resolved only under the highest magnifications of the microscope. Most of the plates are elongated parallel to one pair of the side boundary lines. The maximum crystal length is about 0.015 mm. Many of the plates are loosely stacked in groups of 5–10. Although the crystallinity suggests dickite, an X-ray powder pattern obtained by Mary Mrose, of the Geological Survey, indicates that the mineral is kaolinite.

The mineral clumps, which are sparsely and inconspicuously scattered in the enclosing rock, constitute the only occurrence of crystalline kaolinite recognized in the field. Enough specimen material was collected and examined from localities throughout the field to indicate that it is, at best, a rare constituent of the ores.

Barite ($BaSO_4$)

Barite is of only local occurrence in the Picher field. It is common in caves at the base of L bed in the Dobson mine. There it occurs in thin translucent nearly white to pale-greenish-blue crystalline plates that are compounded into rosettes and radiating coxcomb and sheaf-like masses as much as 3 inches long. These are in part imbedded in botryoidal and reniform masses of calcite, or in rosettes of the flat rhombohedral type of calcite. Some of the barite has been leached to a nearly white opaque honeycombed mass, which is readily broken out, leaving casts of the blades in the various types of calcite.

Small crystals of barite were also found at one locality in the Kenoyer mine (NE¼SW¼ sec. 20). Ac-

cording to Weidman (1932), barite has been found on galena crystals in the Grace Walker and Brewster mines.

Gypsum ($CaSO_4 \cdot 2H_2O$)

Gypsum is usually a weathering product that was deposited after the mines were dewatered. However, a different type occurs at one locality in the Dobson mine and is believed to be primary in the sense that the mineral was crystallized under conditions of complete ground-water saturation, probably at the same general time that the other late gangue minerals were formed. The crystals are clear thick needles as much as ¾ inch long and 1/10 inch in diameter; stockier prisms are ¼ inch long and ⅛ inch thick. These crystals are attached at the base and grow outward into a vug, thus contrasting with the recumbent attitude of the finer grained gypsum found in the oxidized stopes. The associated pink spar, sphalerite, and calcite are fresh and clean, though the surface on some of the calcite crystals is opaque and granulated owing to recrystallization to very fine crystals of gypsum. The common occurrence of barite in the Dobson mine indicates that the sulfate radical was present in the primary mineralizing solutions, and the unique crystals of gypsum are perhaps a further manifestation of this composition.

SECONDARY MINERALS

Minerals formed by oxidation of the primary sulfides are of little importance in the Picher field because most of the ore deposits were below the ground-water table until the mines were dewatered. Only on the fringes of the field, where the ore deposits are at the surface, such as at Lincolnville, were oxidation minerals prominent. These deposits have been largely worked out and abandoned. A little oxidation has occurred since the mines were opened, but the minerals formed are of no commercial importance, although they are of some mineralogic interest.

Smithsonite (*carbonate, dry bone*—$ZnCO_3$; Zn *52 percent*)

The production from the vicinity of Lincolnville from 1907 to 1914 included a little "silicate and carbonate." Although only silicate is specifically mentioned in accounts of the deposits, it is known that this term commonly included the carbonate (Siebenthal, 1908, p. 201). The proportion of smithsonite would appear, however, to have been relatively minor, for whereas carbonate is recorded for adjacent districts, it is not mentioned in Siebenthal's preliminary report on the Lincolnville area (Siebenthal, 1908, p. 205–208). These mines are no longer accessible.

Calamine (*silicate*—$(ZnOH)_2 SiO_3$; Zn *54.2 percent*)

Calamine probably constituted most of the 164 tons of "silicate and carbonate" concentrates that were marketed from the Lincolnville area between 1907 and 1914 inclusive (U.S. Geol. Survey, 1914, p. 107).

Anglesite ($PbSO_4$; Pb *68.3 percent*)

Since the mines of the field have been opened, the galena in many places has developed a thin light-gray coating that resembles putty or wood ash. Chemical tests show this material to be lead sulfate. Most of it is too fine grained to be resolved under the microscope, but locally, there is enough fine-grained crystalline material to confirm the identification through its optical properties. The mineral may also rarely appear as a film of pearly white bladed or feathery scales in cracks in the galena.

Cerussite (*lead carbonate*—$PbCO_3$; Pb *77.5 percent*)

Cerussite is less common in the field than anglesite, but it has been identified in the Anna Beaver and Lawyers mines. At one locality in the Lawyers, minute crystals of cerussite, 1 mm in length, are associated with equally small crystals of gypsum in a small pocket in the base of a galena cube where it fastens onto the underlying material.

At the Zig Zag mine, near Joplin, small well-formed crystals of clear and smoky gray cerussite are found in sheltered recesses and pockets on the surfaces of stalactitic zinc sulfide that contains some galena. Although these sulfides are not obviously oxidized, the associated marcasite has in part altered to an ocherous coating.

Malachite [$Cu_2 (OH)_2 (CO_3)$; Cu, *57.3 percent*]

Minute green spots of malachite that appear on the chalcopyrite and adjacent pink spar in a few mines, particularly the Barr, have obviously been formed by oxidation since the mines were dewatered. The second generation of chalcopyrite that coats and replaces enargite has been especially vulnerable to such alteration. However, the total quantity of malachite so formed is negligible.

Sulfur (*elemental* S)

Very small pale-yellow crystals of sulfur are of common occurrence on the surfaces of oxidizing marcasite. Similar crystals also occur in porous jasperoid from which disseminated sphalerite is being oxidized and leached; they also occur on the surfaces of both fresh and corroded galena. The source of this sulfur is probably oxidation of marcasite and pyrite which are disseminated in the jasperoid or are associated with the

356-188 O—69——9

galena. That elemental sulfur can form at an intermediate stage in the oxidation of pyrite is well known (Lindgren 1933, p. 829), and the chemistry for the oxidation of marcasite should be similar.

Sulfur can also be derived from incomplete oxidation of hydrogen sulfide, which is a common intermediate product resulting from the action of acid mine waters on such sulfides as ZnS. The acidity of the mine waters, however, is ultimately referable to oxidizing marcasite, pyrite, and chalcopyrite.

Goslarite ($ZnSO_4 \cdot 7H_2O$)

Goslarite occurs commonly as a white efflorescence on mine walls and particularly on pillars that contain unmined sphalerite. Part of the goslarite contains varying proportions of magnesium and ferrous iron that are isomorphous with the zinc. The material loses water upon exposure to dry air and recrystallizes to fine-grained products of higher refractive indices, not identifiable with other known minerals.

Epsomite ($MgSO_4 \cdot 7H_2O$)

The occurrence of epsomite is very similar to that of goslarite; the two form an isomorphous series with complete gradation from one to the other. Most of the material from mine stopes has the mixed composition, but relatively pure epsomite is also fairly common where solutions from oxidizing marcasite have acted on dolomite.

Gypsum ($Ca\,SO_4 \cdot 2H_2O$)

Gypsum is a very common product of the oxidation that has taken place since dewatering of the mines. It occurs most commonly as fine needles and feltlike clusters, or as crusts made up of fine rhombic crystals. Because the solutions from which it has crystallized have in general been only films of mine water on the mine walls, the crystals are recumbent on the various surfaces or fill interstices in the more porous jasperoid. Gypsum appears also as microscopic crystals or anhedral blebs replacing gray spar and pink spar dolomite and glauconite, or growing interstitially in porous jasperoid whose crystals it encloses poikilitically.

Melanterite ($FeSO_4 \cdot 7H_2O$)

Melanterite is very common as watery greenish fibrous crusts or as nearly white capillary growths in the vicinity of oxidizing marcasite and pyrite.

Szomolnokite ($FeSO_4 \cdot H_2O$)

Specimens of pyritic jasperoid and massive pyrite, when examined in the laboratory, commonly have a white efflorescence of fine-grained ferrous sulfate. Grains of this material have the refractive indices and the optical elongation of szomolnokite. The material is usually associated with copiapite. The material may have formed in the laboratory from evaporation of moisture contained in the pores of the specimens, but as szomolnokite is commonly formed from highly acid mine waters, at least some of it was probably precipitated underground.

Copiapite [$(Fe, Mg)\ Fe_4\ (SO_4)_6(OH)_2 \cdot 20H_2O$]

Copiapite is a common secondary product of oxidizing marcasite and pyrite, including that contained in porous jasperoid. It is also found on oxidizing chalcopyrite. It appears as minute rough granules or botryoidal growths of yellow, orange, flesh, or gray color, commonly with a translucent waxy or pearly luster. Such growths are composed of fine crystalline aggregates that break into rhomb-shaped cleavage fragments when crushed.

Carphosiderite [$(H_2O)\ Fe_3\ (SO_4)_2\ (OH)_5\ H_2O$]

An earthy sulfur-yellow crust that forms on or near oxidizing marcasite and pyrite is a ferric sulfate mineral, insoluble in water but soluble in acid. It is so extremely fine grained that all the optical properties cannot be determined, but the indices of refraction and birefringence are those of carphosiderite. A considerable percentage of the granules in finely crushed material have positive elongation; these granules are presumably crystals that are tilted up enough to simulate elongation in crystals that are fundamentally basal plates. As jarosite is very similar to carphosiderite in optical properties, qualitative microchemical tests for potassium were made on the material, but proved negative.

Diadochite [$Fe_2\ (PO_4)\ (SO_4)\ (OH) \cdot 5H_2O$]

The crystallized form of diadochite occurs as a pale-pinkish-gray earthy efflorescence at one place in the Central mine. The component crystals are minute rhombic or six-sided plates of characteristic shape, only 0.01–0.02 mm long, and have the optical properties and refractive indices typical of diadochite. The identification was confirmed by positive chemical tests for iron, sulfate, and phosphate. Although formation of the material was doubtless dependent on oxidizing marcasite or pyrite, the acid iron-bearing solutions thus formed may have seeped considerable distances to pick up the phosphate required for diadochite.

PRIMARY CORROSION OF MINERALS

One of the most striking features of the Picher ore deposits is the evidence of intense leaching to which some of the galena was subjected at a temporary stage

during the primary mineralization. Later minerals in the paragenetic sequence were sometimes deposited on the corroded surfaces, but perhaps even more commonly, the later mineral consisted of "resurgent" galena which partly refilled the voids created by the leaching. A similar type of corrosion but without demonstrable resurgence affected sphalerite in places, and less commonly, the chalcopyrite and enargite. Marcasite and pyrite were unaffected. There was also a less conspicuous and local leaching of pink spar crystal surfaces. At one of the few occurrences of barite in the field, the barite has been notably leached.

CORROSION OF GALENA

The common example of leaching is a galena cube containing a rounded cavity particularly on a corner or along one edge, like the cavity in a decaying tooth. A more advanced stage shows an irregularly rounded surface on one or more sides, or perhaps just a globular remnant. If the original crystal had a continuous coating of marcasite, this remains as a residual shell, fresh and untarnished, marking the original boundaries of the cube. Later marcasite may be deposited sparingly on the corroded galena surface. In massive jasperoid, galena may leach out completely or nearly so, leaving a clean cast, or one with a galena remnant, or one partly filled by later minerals, such as sphalerite (fig. 40*A*), calcite, or resurgent galena. Earlier pink spar dolomite, in which the galena may be imbedded, is either not affected by the leaching or only mildly affected; closely associated sphalerite is not affected. Corrosion of an octahedral galena crystal has been noted, but this phenomenon is rare, probably because in most places the octahedral galena is later than the corrosion in the paragenetic sequence.

The corrosion was commonly preceded or accompanied by deposition of an extremely thin film of microcrystalline quartz on the galena surface, like jasperoid except that it has not replaced any preexisting mineral. Such films, left as paper-thin shells upon leaching, may mold part of an original face of the cube, or may record an intermediate stage in which the galena was partly leached. The quartz film is partial to the galena, for closely associated minerals known to be earlier in the depositional sequence, such as sphalerite and pink spar dolomite, are not coated.

The galena surface left by the corrosion may be smooth, but it is more commonly hackly or spongy because of small irregular platy and also wirelike residuals. The plates may be single or multiple, and are most commonly parallel to the cleavage of the original galena, though some are roughly in the cube-diagonal position (fig. 40*B*). The residual "wires," representing a more advanced stage in the leaching, have similar orientations. If later minerals, such as sphalerite, have precipitated on the surface in solid enough mass to embed these residuals firmly, a polished section shows that even the thinnest films are unaltered galena (fig. 40*C*).

Resurgent galena, growing in the space from which the earlier cube was leached, is usually made up of many parallel subcrystals joined together on a common base. Both cubic and octahedral faces are represented in the regrowth, but the octahedron is commonly more prominent than in the original crystal. If the regrowth has taken place on a remnant of the original crystal, the resurgent material is usually oriented so that it is crystallographically parallel to the original crystal. Where the leaching was particuarly severe on an originally large crystal, leaving only small isolated residuals as nuclei on the base rock, the regrowth can give rise to many small crystals which, though isolated from each other, are in perfect parallelism or nearparallelism (fig. 40D).

Leached galena is widely distributed in all parts of the field, though it is sparse in the southwestern area where jasperoid is the prevailing gangue to the near exclusion of gray spar dolomite. In the more dolomitic areas, leached galena is equally common in the dolomite blocks, in the jasperoid walls near such blocks, and in adjacent jasperoid areas less closely related to the dolomite.

Leaching of galena from beneath fresh marcasite and pyrite crusts was also noted in the Joplin district by Siebenthal (1915, p. 45, pl. 5), who ascribed the phenomenon to differential oxidation and removal of the sulfide with lesser electrical conductivity where in contact with one of greater conductivity. In the Picher field, however, the leaching has taken place in unoxidized ground. Similar leaching of galena, with no affect on associated pyrite and chalcopyrite, has been described in a hydrothermal vein near Central City, Colo. (Bastin and Hill, 1917, p. 136–137, 212). Bastin concludes that the leaching was effected by a late phase of the hydrothermal solutions which deposited quartz and siderite gangue in adjacent parts of the vein.

CORROSION OF SPHALERITE

Leached sphalerite, though widely distributed, is neither so common nor so conspicuous as the leached galena. Its distribution is entirely independent of the leached galena, so much so, that occurrence of the two at the same locality is rare. The leaching of sphalerite shows no prevalence in any one type of ground environ-

A. Cast of leached galena crystal in jasperoid, the cast containing a globular remnant of corroded galena hanging from the roof and later crystals of ruby jack (not visible in the photograph); from Lucky Syndicate mine.

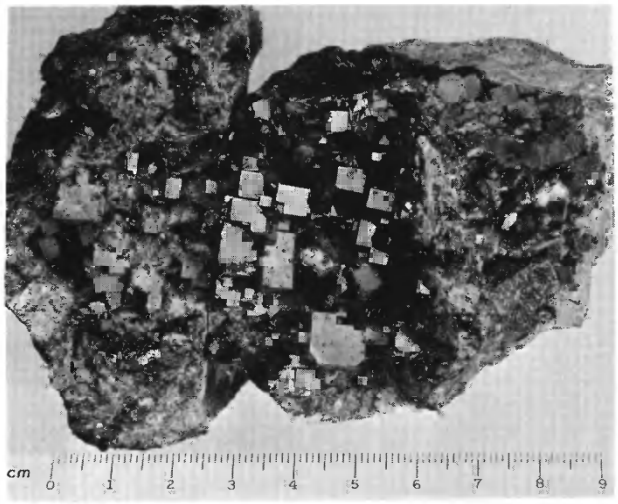
B. Leached galena showing thin residual plates mostly parallel to cleavage, but some in cube diagonal position; from Barr mine.

FIGURE 40.—GALENA SPECIMENS FROM PICHER

ment over another, but there is a marked tendency for it to affect ruby and rosin jack rather than black jack. The interior of a crystal is more vulnerable than the surface, and in many examples the center of the crystal leached out, leaving an exterior shell. In some places the crystal is coated by marcasite, pyrite, or chalcopyrite, which is left intact in the residual shell. Particularly striking are globular shells of marcasite found at one locality in the Kenoyer mine. Whereas some of the marcasite shells contain a sphalerite core, others have had the sphalerite removed, leaving an empty shell (fig. 32). Where a sphalerite residual remains, the surface of leaching is usualy deeply and irregularly fluted by the solution, and shows a bright shiny reflection. The leaching has in places been very intense without affecting closely adjacent pink spar dolomite.

Evidence dating the corrosion as part of the primary mineralization is not so convincing as it is for galena. Deposition of small octahedral galena crystals on the corroded surfaces has taken place at one locality in the Blue Bonnet mine, and at a few other places fine pyrite or marcasite has a similar occurrence, or calcite has crystallized in the space leached behind a marcasite shell. Rarely, resurgent sphalerite crystallizes in small (an eighth of an inch) and poorly formed tetrahedrons in parallel orientation, loosely bound together in the

C. Polished section showing thin films left by leaching of galena, which are imbedded in a later crystal of sphalerite; from Midas mine. The film on left was originally at the cube surface of a galena crystal, of which only a small sector bordering the corrosion cavity is here shown. Galena, gl; sphalerite, sl; cavity, c.



D. Galena crystals in parallel orientation owing to resurgent growth in an old cast of galena that contained enough solution residuals to control orientation of the second generation; from Goodwin mine. Imperfections in parallelism are no greater than in different parts of a typical large unleached crystal.

FIELD SHOWING CORROSION PHENOMENA

space from which earlier sphalerite was leached, particularly just under the unleached surface shells of large crystals. Such resurgent sphalerite may have later small crystals of marcasite growing on it.

CORROSION OF CHALCOPYRITE

Conspicuous leaching of the chalcopyrite is widespread. The leaching has not noticeably affected sphalerite in which the chalcopyrite may be imbedded, nor has it affected marcasite and pyrite which are imbedded in or grow on the chalcopyrite crystals. Inclusions of iron sulfide minerals are laid bare in the leaching, which is particularly concentrated adjacent to these inclusions; and coatings, particularly of pyrite, are left as a residual shell. Usually, the leaching exposes the fresh chalcopyrite in an irregularly pitted surface, though locally, the pitting on a sphenoid face may be triangular or terraced parallel to the face edges. The corrosion antedated deposition of some sulfides, for in polished section (Barr Mine), octahedral galena covers and fills some of the irregularities in the pitted surface. There is no correlation with similar leaching of galena and sphalerite, for although these may be leached in the same general areas, rarely is more than one sulfide leached at a given locality. Primary leaching is distin-

guished from recent oxidizing alteration by the absence of alteration products, such as iron oxide and malachite.

Corrosion of sphenoidal chalcopyrite accompanies the replacement of enargite by a later botryoidal form of chalcopyrite at one locality in the Grace Walker mine, and the suggestion is made (p. 110) that the two processes are complementary and related.

CORROSION OF DOLOMITE

Locally, the pink spar dolomite in vugs shows superficial leaching, which consists of fine pitting or scaling of the crystal surfaces, accompanied by bleaching of the pink color. This phenomenon was noted chiefly in several mines in Kansas near the State line—the Barr, Webber, Bendelari, Midcontinent, King Brand, and Boska—but was also found at the Dobson in Oklahoma. It evidently was earlier than the opening of the mines, for the calcite that is present with leached dolomite in places is unleached, and associated marcasite is unoxidized. The finely crystalline marcasite (prismatic to hair type) that usually accompanies this leached dolomite is either fresh or only moderately tarnished, and at first sight it appears to be earlier than the leaching in that it occurs only on remnants of the original crystal surfaces of the dolomite. However, the selectiveness of marcasite for the distal edges of pink spar crystals as sites for crystallization (p. 113) would doubtless prevent its growth in the leach pits; hence, the leaching may possibly have been contemporaneous with or earlier than the marcasite. Where chalcopyrite is coated over the corroded pink dolomite crystals, it fills the small leach cavities in their surfaces; but the relations are indistinguishable from those interpreted as replacement of the dolomite by the chalcopyrite (p. 108). There is no consistent association of leached dolomite with previously described leaching of the sulfides, though there may be uncommon and apparently random association with the leached galena or chalcopyrite.

CORROSION OF BARITE

Much of the barite in the Dobson mine has been intensely leached to white opaque honey-combed masses. This leaching antedated deposition of the botryoidal calcite in which the barite blades are imbedded. The fine interstitial cavities resulting from the corrosion contain a few minute granules of gypsum.

PARAGENESIS

Paragenetic relations of the minerals in the Picher deposits are of particular interest for the clues that they may yield on the origin and conditions of deposition of these ores. The general occurrence of caves and vugs in the ore deposits has permitted superposition of different minerals in the order in which they crystallized. Interpretation of this order is complicated by affinities and antipathies between certain minerals, the causes of which are not understood. Sequence of mineralization is also indicated by veining of earlier materials by later ones, though this criterion has to be used with considerably more caution because of possible late replacement in the mass that borders veins.

The generalized order of crystallization and the numerous overlaps in this order are indicated in figure 41. The initial step in the mineralization was replacement of part of the limestone country rock by dolomite to form massive gray spar, with pink spar along the more open bedding seams, veinlets, and vugs. Although most dolomite was deposited before the jasperoid and sulfides, some was recrystallized to pink spar during the jasperoid stage, and in rare instances, crystals of pink spar are found on sphalerite or cubic galena crystals.

The dolomite was followed by massive jasperoid that replaced the gray spar to some extent, but also spread far beyond the confines of the dolomitized ground to replace extensive additional masses of limestone. The original limestone was much more vulnerable than the dolomite to the jasperiod, and what in many underground exposures appears to be replacement of dolomite, particularly along bedding bands, actually is replacement of residual limestone that had escaped the initial dolomitization (fig. 42; McKnight, 1935, p. 112, 119). In general, the replacement of limestone was by massive pervasion of the silica-bearing solutions, like water into a sponge, and any lack of completeness is expressed texturally in the amount of calcite remaining interstitial to the microscopic jasperoid grains. In contrast, where dolomite is replaced by jasperoid along fractures transverse to the bedding (which is the only case for which replacement of the dolomite can be ascribed with certainty), the replacement is largely confined to a comparatively narrow band along which the action was concentrated, and the lack of completeness is obvious in the macrotexture of the resultant heterogeneous pattern. Along contacts between jasperoid and overlying gray spar, the dolomite was commonly recrystallized to pink spar, apparently in space produced by contraction of the jasperoidized mass (p. 121).

The jasperoid deposition overlapped that of sphalerite and galena, which are later in the generalized sequence. As jasperoid is largely restricted in its occurrence to massive replacement of limestone and dolomite, criteria for judging its age relative to the sulfides are few. Minute crystals or, more rarely, continuous films of

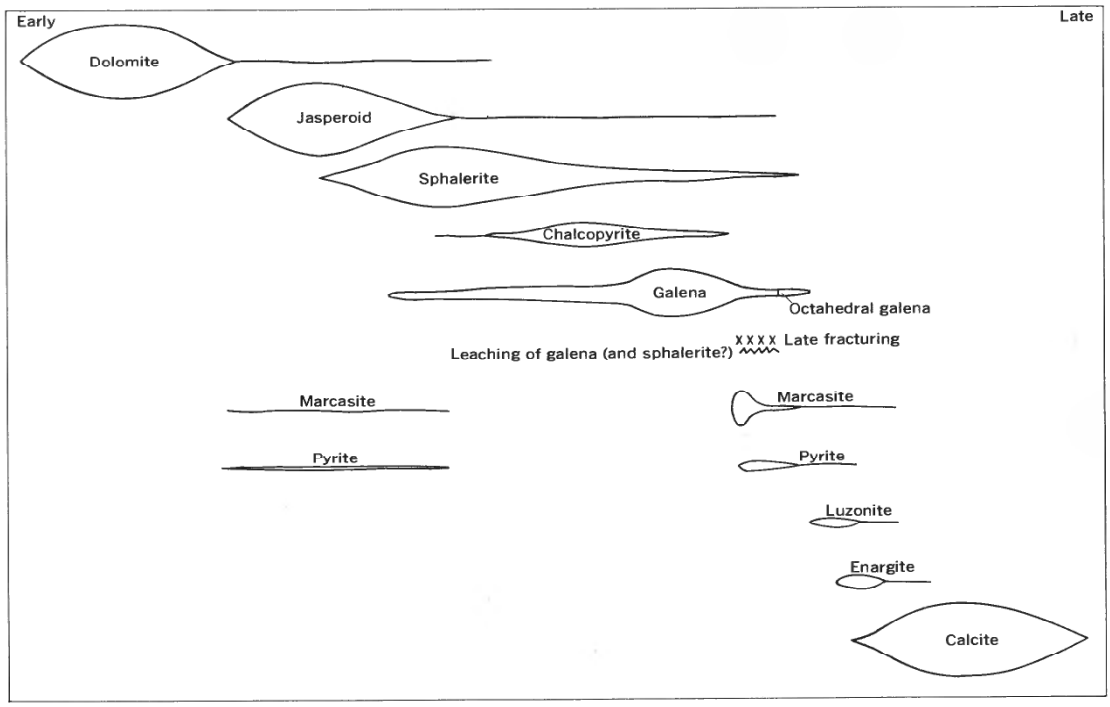


FIGURE 41.—Generalized order of crystallization of the primary minerals in the Picher field. (Width of bleb is a diagrammatic indication of the mass precipitating at different stages during the mineralization; the length does not represent continuous crystallization, but rather the time interval during which intermittent precipitation took place at varying and indefinite periods.)

microscopic quartz are occasionally found on the contact between pink spar and sphalerite from vugs, and presumably were contemporaneous with deposition of part of the jasperoid. Minute crystal faces of quartz on the free surfaces of jasperoid in vugs are, however, only in part earlier than vug sphalerite, for in many places they do not extend underneath adjacent sphalerite crystals. The almost universal occurrence of jasperoid in the matrix of disseminated sulfides suggests that silica was still being deposited from the solutions at the time of the ore-depositing stage. At some localities on the borders of mineralized ground, however, the limestone enclosing disseminated sphalerite and galena is only weakly silicified, and exceptionally, no disseminated jasperoid can be found in thin sections of the mineralized limestone. During the main surge of zinc-lead mineralization, most of the sphalerite and galena was deposited along or adjacent to the more open channelways and solution caves without accompanying quartz, and it is probable that the silica content of the solutions may have been greatly diminished by the time this stage was reached. However, some silica persisted virtually to the end of the zinc-lead deposition, for it appears in a few places as a quartz coating on vug sphalerite and galena, or more commonly, as a jasperoid film on crystal faces of cubic galena. At one locality in the Goodwin mine, an incomplete jasperoid film is enclosed in a galena cube parallel to, and a short distance in from, two of its crystal faces. A polished cross section shows that this film is on an obscure but definite surface that marks a late pause in the growth of the crystal. Practically all late jasperoid films on galena are accompanied by the primary corrosion of the galena previously described (p. 125).

Sphalerite, chalcopyrite, and galena were deposited in general in the order named, but there are numerous localities that show reversal from this order. Sphalerite replaced the gray spar extensively, galena replaced it only rarely, and chalcopyrite shows no clear-cut replacement, though it replaces pink spar to a minor extent in vugs. Relative order between the three sulfides is revealed in the open cavities where the earliest is generally sphalerite, usually but not invariably in coarse crystals. Late sphalerite is in small crystals, rarely as much as three-quarters of an inch across, that may be perched on the chalcopyrite, on the cubic galena, or on leached surfaces of galena. The reversal from the "normal" order of crystallization between sphalerite and galena is not entirely haphazard in its distribution, for

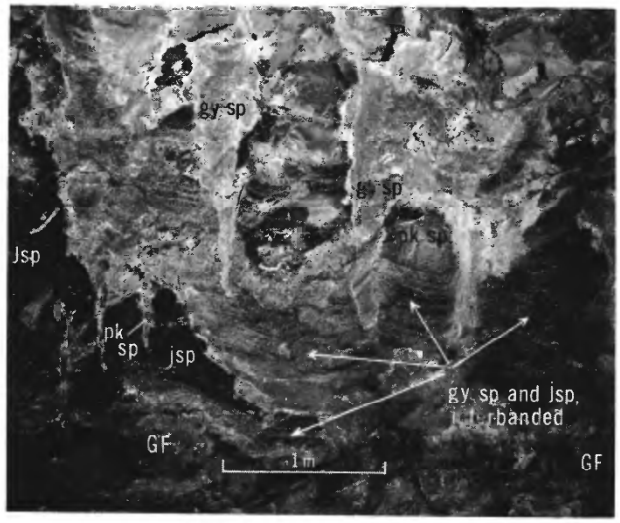


FIGURE 42.—Dolomite and jasperiod replacing M bed limestone, New Chicago No. 1 mine. The pendant features, resembling "stalactites," are gray spar (with a bordering fringe of pink spar) which formed in the limestone along near-vertical fractures; the gray spar banded parallel to bedding formed at the same time. At a later stage, the residual limestone, including that interbanded with the gray spar, was replaced by jasperoid. Gray spar, gy sp; pink spar, pk sp; jasperiod, jsp; Grand Falls Chert, GF.

it shows a zonal relation to ground in which the normal order prevails, as is pointed out in the discussion of mineralogic zoning.

Chalcopyrite is characteristically on the sphalerite or on the pink spar dolomite that preceded the sphalerite. It may, however, be enclosed in sphalerite crystals in a concentric zone that is usually near the periphery, or it may underlie late sphalerite crystals where it was obviously earlier. In a much less common association where chalcopyrite is in sequence with galena, it usually underlies the galena, but, rarely, it is enclosed in galena cubes or is perched on their surfaces. Octahedral galena is always later than cubic galena, and usually later than associated sphalerite and chalcopyrite, though exceptionally, it has sparse small crystals of sphalerite or chalcopyrite perched on it, or has crystals of chalcopyrite included within it.

Although it might at first appear that these three sulfides were deposited more or less contemporaneously, the peak of deposition for each coming at a different stage in the precipitation interval, this is not the true situation. At any given locality the vug minerals are in definite generations, and in general, the earlier mineral had practically completed its crystallization before the one following started to precipitate, with the chief exception noted above wherein chalcopyrite is enclosed zonally in other sulfide crystals. Sphalerite and galena crystals grow on each other, which shows that there is no mutual exclusion of either one toward the other. Although it is conceivable that molecular forces during slow simultaneous crystallization of these two minerals could prevent intergrowths or random inclusions of one mineral in the other, mutual overgrowths should be expectable under such conditions. Usually, however, the sequence at a given locality is in one direction only; and if there are inversions, they are clearly ascribable to multiple generations.

Not uncommonly there are, at a given locality, two generations of sphalerite with cubic galena between them. The late sphalerite crystals are smaller than those of the earlier generation, but in the absence of galena or some similar dividing mass, the later zinc sulfide is also deposited as a surficial layer on the earlier and is usually identical with and indistinguishable from it. The earlier generation was crystallized before the galena was deposited, though locally, there may be a slight penetration of its surface by the galena base. It is probable that at the time the later small sphalerite crystals were deposited on the galena, an accretion layer was added to the earlier sphalerite, thus accounting for the implantation of the galena root. That the growth of the earlier sphalerite phase may have been pulsating is further shown by the chalcopyrite crystals along concentric zones in the sphalerite crystals. Exceptionally, two or three such zones may be present, each recording enough of a pause to allow growth of the chalcopyrite, which, however, is always in small grains where so included. The composition of the depositing solutions may have changed locally, and the outer part of sphalerite crystals may show an abrupt change in composition and color from that in the core (p. 103). The very latest sphalerite crystals, as distinct from accretion on earlier crystals, are those few, always small, that are found on octahedral galena and on marcasite.

The chalcopyrite was most obviously periodic in its crystallization where it lies at certain zones in the sphalerite, but it is formed at several other stages, as previously indicated. A very late generation occurs exceptionally at a definite zone in certain large calcite crystals, generally at the boundary between the purple core and the more nearly colorless peripheral shell.

Severe leaching separated the galena into at least two recognizable generations of cubic habit, and there is also the later octahedral phase. Generally, only two phases are found together; at the one locality where perhaps all three phases were observed, superposition of the latter two, though probable, is not definitely proven. Sphalerite crystals exceptionally as much as three-

quarters of an inch across may lie on the corroded galena surface between the two generations of cubic galena, or smaller crystals may separate earlier cubic galena from later octahedral galena. Such sphalerite is also common in galena casts where the galena has been leached from jasperoid (fig. 40*A*).

A period of minor fracturing was approximately contemporaneous with leaching of the galena and, indeed, may well have initiated the chain of events that led directly to the leaching.

An early massive form of pyrite replaced chert and limestone, and was probably earlier than the ore mineralization. On free edges, bordering openings, it was locally altered to marcasite, probably during the ore deposition. Microscopic pyrite and marcasite that occur in jasperiod are apparently contemporaneous with the jasperiod. Most of the marcasite and at least the conspicuous part of the pyrite, however, are relatively late minerals in vugs, later than the main surge of cubic galena on which they are most prominently encrusted. They are also common on pink spar, sphalerite, and chalcopyrite, the last of which they replace extensively. The main period of deposition was rather sharply limited to a stage following most of the cubic galena but apparently preceding the octahedral galena, for the latter not only overlies marcasite and pyrite, but elsewhere has only minor incrustation by scattered small (usually minute) crystals of these iron sulfides that represent the waning stages of their deposition. In places, marcasite and pyrite lie on the surface of galena cubes, but beneath late octahedral projections that grow on such cubic surfaces. The marcasite and pyrite were largely earlier than the corrosion of the galena; but they were also deposited on the corroded surfaces, though nowhere so abundantly nor in such coarse crystals as those on the unleached surfaces. The final iron sulfides were the very small crystals that occur in zones in some of the large calcite crystals. Whereas the pyrite in jasperoid is conspicuously octahedral, that in the later vugs is cubic or pyritohedral, or some combination of these two.

The two iron sulfides are probably virtualy contemporaneous. In places, small pyrite crystals are on marcasite crystals, and hence are later. In polished sections, however, it is common to find pyrite crystals engulfed in the base of marcasite crystals, or the two minerals may be intergrown anhedrally for some distance above the base, indicating the development of some pyrite earlier than marcasite or during the early stage of marcasite growth.

The primary leaching of sphalerite occurred at about the same time as that of galena. As it is not so common, its period cannot be delimited so closely, but it at least followed deposition of marcasite and pyrite crusts, preceded late fine crystals of the same minerals and of octahedral galena, and definitely preceded the calcite. Leaching of pink spar and of chalcopyrite, observed at a few localities, also occurred at approximately the same stage; thus, at one locality in the Barr mine, octahedral galena covers and fills corrosion cavities in chalcopyrite. However, the leaching of these various minerals was not coextensive in space, for there are numerous examples wherein the galena is completely dissolved, whereas early sphalerite or pink spar in contact with it are unaffected; the leaching of chalcopyrite commonly has taken place without affecting the adjacent sphalerite.

Crystals of enargite and the crystallites of luzonite are common on pink spar, sphalerite, and particularly chalcopyrite. Although later than galena, they have never been observed on it, but they appear on chalcopyrite that is on galena. They likewise avoid marcasite and pyrite even though all may be common in the same vug. In the few places where age relations have been determined, enargite and luzonite are later than marcasite; this sequence is also in harmony with the conclusions that might be derived from observing their mutual relation to calcite. In general, all precede the abundant calcite; but whereas only small and insignificant crystals of marcasite overlap and are included zonally in the calcite crystals, well-formed coarse enargite crystals with accompanying luzonite may be included in calcite or even be perched on the surface of large scalenohedrons.

In the only association with pyrite in which the relative ages could be determined (Barbara J mine), enargite is later than and molded around a small crystal of pyrite (cube + pyritohedron), but, in turn, it has still smaller pyrite (pyritohedron, 0.15 mm in diameter) on its crystal surfaces.

The occurrence of microscopic enargite blebs in chalcopyrite crystals, as revealed in polished section, possibly represents replacement virtually contemporaneous with deposition of enargite crystals on the chalcopyrite surface. On the other hand, the common association of miscroscopic bleb enargite with marcasite and pyrite that have replaced chalcopyrite may indicate contemporaneity with these iron sulfides; some of the copper removed in replacing $CuFeS_2$ by FeS_2 might well be trapped in the immediate environs in the iron-free Cu_3AsS_4. The relations are far from quantitative, for the pyritic-marcasitic replacement is abundant, whereas the enargite association is rare. It is also

possible that some blebs of enargite in chalcopyrite may represent primary intergrowth.

The luzonite crystallites are approximately contemporaneous with the enargite. Usually, the two grow from some common base so that there is no evident difference in age; but specimens have been observed in which crystals of enargite grow on the basal arms of luzonite crystallites and are thus later. The reverse order is rare, but it does occur.

The replacement of enargite by a massive (finely botryoidal) form of chalcopyrite, with recrystallization of some of the enargite to more delicate crystals and to slender delicate crystallites of luzonite (p. 110), is a late phenomenon which, nevertheless, preceded crystallization of the flat rhombehedral form of calcite in at least one locality (John Beaver mine). Late chalcopyrite does not everywhere replace the enargite. Thus, figure 43 shows a polished section of a specimen from the Admiralty mine, in which microscopic late chalcopyrite replaces pink spar below a thin selvage zone of enargite which undoubtedly marks the original base of a large chalcopyrite crystal. Such late chalcopyrite is, however, exceptional other than in the crudely pseudomorphic replacement of detached enargite crystals (p. 108, 110).

Calcite is in most places the latest mineral to have crystallized. Locally, chalcopyrite, marcasite, or enargite may be enclosed zonally in it, but such occurrences are not common, and the first two have been observed only in calcite caves near, but not actually in, ore-bearing ground. The relative ages of the different types of calcite have been discussed in the section on mineralogy of calcite (p. 119); the general order, from oldest to youngest, seems to be scalenohedral, prismatic, and flat rhombohedral.

The sparse barite is later than marcasite, and earlier than the botryoidal calcite with which it is associated in the Dobson mine. It was severely leached before deposition of the calcite.

MINERAL ASSOCIATIONS

A notable feature of the mineralogy in the Picher field is the association of certain minerals, such that a later mineral is dependent on the environment of one formed earlier for conditions favorable for its crystallization. A related phenomenon is the reverse relationship, in which a later mineral avoids contact with certain earlier ones.

The most persistent association is of chalcopyrite with dolomite. With few exceptions, the chalcopyrite is confined to blocks of ground that contain the gray spar replacement of the original limestone. There may be occurrence in closely adjacent ground, but in such cases

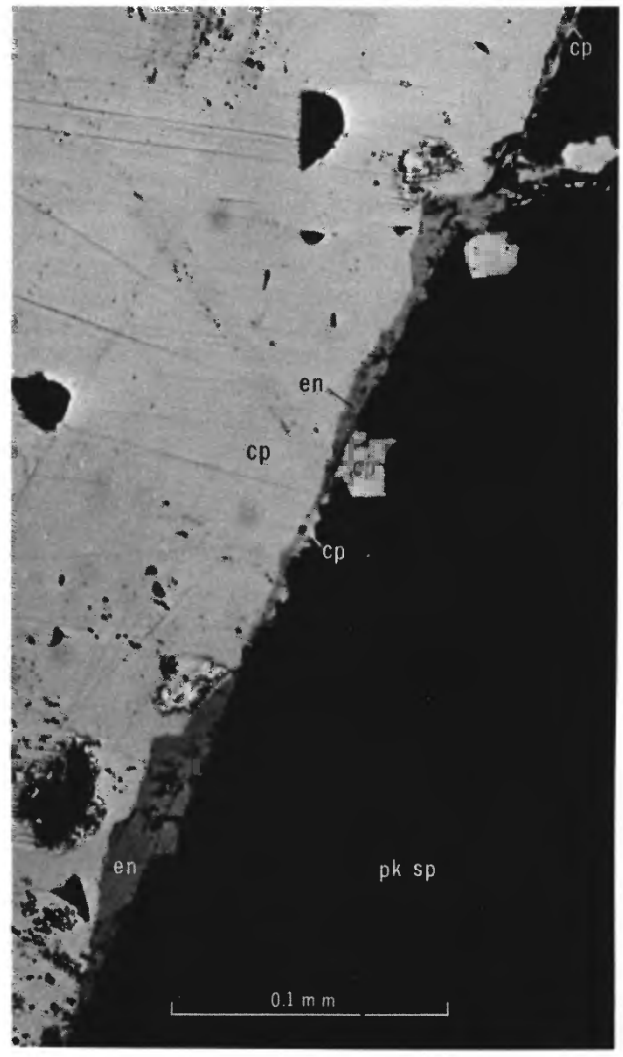


FIGURE. 43.—Photomicrograph of polished section showing enargite at contact of chalcopyrite crystal with adjacent pink spar, from Admiralty mine. The enargite evidently replaces the edge of the chalcopyrite crystal at the contact of chalcopyrite with pink spar. Note that boundary of enargite with later chalcopyrite (to right of it in photograph) is relict from the original chalcopyrite-pink spar boundary. Enargite, en; chalcopyrite, cp; pink spar, pk sp.

pink spar dolomite, which overlaps to some extent the boundaries of massive replacement blocks, is usually present. The chalcopyrite is not necessarily deposited in contact with dolomite, but the dolomite environment appears to be essential. Only the relatively minor occurrence of chalcopyrite in the large barren calcite caves fails to show the dolomite association.

The rarity with which chalcopyrite is found growing on galena is in part due to the generally earlier appear-

ance of chalcopyrite in the paragenetic sequence. There is, however, apparently an additional factor—a certain antipathy of the chalcopyrite toward earlier formed galena. Specimens are occasionally found in which numerous crystals of chalcopyrite that appear to be on galena are actually perched on single small crystals of sphalerite; the sphalerite crystal forms a nucleus for the enveloping chalcopyrite or may even hold it from contact with the underlying galena.

Enargite and luzonite closely follow the distribution of chalcopyrite, and thus, indirectly, they also are dependent on dolomitized terrain. Within the chalcopyrite environment, they may occur on the surface of sphalerite crystals or on pink spar dolomite, as well as on the chalcopyrite; enargite may also replace chalcopyrite internally. Enargite and luzonite are not found on galena even though they are always later than the galena.

The form of sphalerite known as black jack, which has a distinctive habit of crystallization (p. 102), is associated rather closely with dolomitized terrain. Black jack occurs either in a block of ground in which the limestone has been replaced by gray spar, or else in close proximity (within a few tens of feet) to such ground. Thus, it has approximately the same distribution as the pink spar. Rosin jack and ruby jack may also be found in the same association with dolomite, but they are more characteristic of jasperoid-bearing ground. Generally, the color type of sphalerite is uniform at a given locality, but finely crystalline late phases of rosin and ruby jack may be associated with all other types, including the black jack.

Marcasite has widely variable associations, but the bulk of it occurs in openings in jasperoid ground rather than in dolomite, even though small crystals are common in the dolomitized ground. Marcasite is strongly attracted by cubic galena, and examples are common in which the galena is extensively covered with a marcasite coating, whereas other minerals that are also earlier than the marcasite are only sparingly coated or are free of any marcasite. Pyrite, which is a much less common vug mineral, has the same partiality for galena as the marcasite. Both these minerals are also more likely to occur on other sulfides, such as chalcopyrite and sphalerite, than on earlier nonsulfide minerals. There are exceptions, however, such as at localities in the Mary Jane and Anna Beaver mines where fine marcasite is abundant on pink spar, especially on the distal ridges of the crystals, and also on adjacent jasperoid, but is only sparingly present on closely associated chalcopyrite. Perhaps the chalcopyrite is formed later than the marcasite in these places, but such a sequence is at variance with the usual order of crystallization. Marcasite and chalcopyrite are deposited on pink spar and avoid the equally accessible chert along a late fracture in the New Chicago No. 2 mine.

The relatively late jasperoid that accompanied corrosion of galena was deposited commonly as a thin film on the corroded surfaces as well as on the original cube faces, but it generally did not coat any closely associated sphalerite crystals that preceded it in the sequence of crystallization.

Calcite is a widely distributed late mineral which, although common in dolomitic ground, tends to have a greater bulk in jasperoid ground, possibly because there were more voids here at the time of calcite deposition.

SIGNIFICANCE OF PARAGENETIC RELATIONS

Paragenetic relations show a sequence of changes in mineral deposition from the earliest dolomite to the latest calcite. Nevertheless, within this sequence, in places some minerals crystallize in a reverse order. This reversal is readily explainable by the overlap in the time spans during which the different minerals were deposited. Late generations of the several minerals, most readily recognized in the reverse sequences, are in smaller crystals and smaller total mass than the main surges of those minerals (fig. 41). The total picture presents a secular pattern in which each mineral precipitates most of its bulk at a certain stage but continues to precipitate smaller quantities at later brief and sharply defined stages, trailing off into insignificance at the more recent end. Thus, the marcasite and pyrite crystals that formed on leached or octahedral galena are smaller and sparser than the earlier material on the cubic galena; the chalcopyrite crystals that formed in the casts from which galena has been leached, or inside calcite crystals, are smaller than most of those occurring on pink spar or sphalerite. Minerals formed on the fractures that developed toward the end of the ore deposition period are smaller than most crystals of earlier generations of the same minerals. The latest sphalerite to crystallize is in fine well-formed crystals of the several color types, including the black jack. Some observers (Smith, 1935) have interpreted it as a secondary form of zinc sulfide. However, it shows all gradations in size and color to the coarser sphalerite with which it may be intimately associated in the ore bodies, and there is no reason to believe that it represents other than the final crystallization in the dying stages of the primary sphalerite mineralization. This does not necessarily mean that the finely crystalline material is everywhere contemporaneous, for accidents of material supply during earlier phases of crystallization may have stunted the growth of relatively early crystals. The

fine rosin or ruby jack phase is commonly the only sphalerite present on the borders of the ore bodies, but it was probably contemporaneous with the main sphalerite surge in the ore bodies.

ASYMMETRIC OVERGROWTHS

A common phenomenon noted in specimens that have later minerals superposed on cubic galena is the relative abundance of the later mineral, particularly marcasite, pyrite, and fine-grained sphalerite, on certain of the cube faces, and complete or relative absence on other faces. The favored surfaces are those showing a component of face in a certain direction, and the relatively free surfaces are those that face in the opposite direction. Similar relations have been observed where fine quartz is coated on sphalerite. One interpretation is that during deposition of the overgrowing mineral, the mineralizing solutions were moving slowly through the ground and thus carried in solution the substance to be precipitated to the stoss side of the host mineral where most of the solute was deposited (Newhouse, 1941, p. 620, 623, 627; Stoiber, 1946).

Where asymmetric overgrowths on galena were observed in place by one of the authors, the overgrowths were on the upward faces of crystals, and all were close under a solid ceiling that would have precluded any general movement of the depositing solutions downward. The controlling factor in such occurrences is obviously gravity, which probably acts on the seed crystals to hold them in place on the upward-facing surfaces of the host mineral, perhaps after a short settling through the solution. Furthermore, at the time the overgrowths were deposited the solutions must have been virtually stagnant.

PARAGENETIC RELATIONS IN ZIG ZAG DEPOSIT

Mineralization at the Zig Zag mine began with replacement of the limestone by massive gray spar dolomite, containing vugs of the pink spar phase. The earliest sulfides are the sphalerite and wurtzite composing the stalactites. They are so intimately intermixed as to suggest their contemporaneity throughout the crystallization of the stalactites. However, near the end, only wurtzite was deposited in well-formed crystals on the stalactite surfaces. Galena formed radial blades extending from near the centers of the stalactites to octahedral crystalline terminations on their surfaces, and thus it was approximately contemporaneous with the zinc sulfides. It also may form boxworks of minute blebs in the zinc sulfides, elongated and alined in two intersecting directions as seen in polished section, which may be related to cubic directions of growth. Pyrite and marcasite in general occur on the surfaces of the stalactites or fill the interstices between the stalactites and enclose the wurtzite crystals. They are, therefore, in general later. In one polished section, the marcasite of the interstitial material is greatly subordinate to the pyrite in which it is embedded, and is localized in the immediate vicinity of the wurtzite. In places, pyrite or marcasite also may occur with galena in radial blades in the outer shells of stalactites. A colliform type of pyrite is in places interlayered with outer shells of the zinc sulfides. Latest to crystallize were the small crystals of golden-yellow sphalerite peculiar to this deposit.

STRATIGRAPHIC DISTRIBUTION OF THE ORES

All strata of the Boone Formation from the Reeds Spring Member to the top have been mineralized to some extent at some place or other within the Picher field. In addition, strata of the Chester Series have been richly mineralized in many places, but mostly in the lower part adjacent to the Boone, or where they are relatively thin. Because of the unconformity at the base of the Krebs Group, the thickness of the Chester strata preserved beneath it varies abruptly from place to place, and in some of the thicker local sections there may be strata that are nowhere mineralized. As the Quapaw Limestone has been only imperfectly differentiated from basal strata of the Chester Series in the mining field, possibly some of the mined ore classed as Chester may have come from this limestone. Although the Krebs Group contains no ore in the Picher field, its basal strata locally contain a little sphalerite in thin veinlets and pockets, particularly where the subjacent strata are mineralized. Possibly such traces of mineral are common higher in the formation, but are not seen for lack of mine workings in these parts of the section, though traces of mineral are revealed by drilling in some places, as in the Warner Sandsone Member on the Thompson tract (NE¼SE¼ sec. 36, T. 34 S., R. 23 E., Cherokee County, Kans.). Shale and sandy seams of Krebs age occurring as cave filling in the Boone Formation commonly contain traces of mineral comparable to those in the basal part of the shale at higher levels; but such material is not favorable for any significant ore and may even, because of its imperviousness, insolate blocks of limestone that might otherwise have been mineralized. In the Missouri part of the Tri-State region, particularly north of Joplin and in the Neck City area still farther north, the lower part of th Krebs Group shale where it has been broken by slumpage into underlying sinkholes in the Boone has in places been mineralized to the extent that it could be mined by opencut methods in the early 1950's. The Snapp property, a few miles

north of Joplin, is a typical example (K. L. Cook, unpub. data, 1954).

Even though all horizons of the Boone may be locally mineralized, certain beds have proven to be far more productive of ore than others. The Joplin Member (virtually M bed of Fowler) has been by far the greatest producer. Other particularly favorable stratigraphic zones for ore are K bed, the G-H bed interval, E bed, and parts of the Grand Falls Chert Member except N bed (table 2, p. 20). All these beds consist of chert in limestone. The chert gives an overall brittleness to the stratigraphic unit which is expressed in widespread shattering in response to slight preore structural stress, and the limestone furnishes a medium which is easily replaced by the ore and gangue minerals. The importance of rock breakage under such conditions as to create the greatest degree of permeability in voids greater than those involved in the grain porosity of the host rock has been stressed by Rove (1947, p. 186–192) as a necessary prerequisite to ore deposition.

In M and K beds, the optimum ratio of chert to limestone is apparently reached, and these beds have yielded some of the richest ore in the field. K bed has been less productive only because its favorable eastern facies is less extensive areally. The fact that both of these beds have massive chert beds under them, M even lying between these two cherts, may also be an important factor in that such relatively unyielding strata must concentrate into the overlying ore zones a maximum breakage of the rock caused by relatively slight warpings of the strata.

A massive chert, such as that in the upper part (N bed) of the Grand Falls Chert Member, evidently did not yield by well-distributed shattering; furthermore, it lacked the chemical qualities to make it a good precipitant of the ore minerals. The sheet-ground zone of the Grand Falls Chert Member lies 20–30 feet below the base of M bed and is enough isolated from M bed ore that separate workings have to be made into it to recover the ore; but although dominantly a bedded chert zone, it originally contained enough interbedded limestone, perhaps 10 percent, to facilitate widespread shattering and precipitate such quantity of ore minerals as to constitute low-grade ore of wide and continuous distribution. This ore is profitably minable by virtue of its extent, uniform thickness, and general horizontality, all of which make for low mining costs. Sheet-ground ore is usually not more than 10–12 feet thick. It is chiefly localized in a few parts of the field, notably the southwestern part in contiguous workings of the See Sah, Blue Goose No. 2, Humbahwahtah, and South Side tracts, and the northeastern part of the field in contiguous workings of the Bailey, Ballard, Shanks, Hartley, Liza Jane, Leopard and other tracts; but there are several smaller areas northwest of the latter group in Kansas. As most of the sheet-ground deposits were opened after fieldwork for the present report was terminated, they have been relatively little studied.

The basal 6–8 feet of L bed chert differs from that in the remainder of L bed and in the barren parts of the Grand Falls Member in being more thinly stratified, so that locally it gave a good shatter pattern in response to preore stresses that were otherwise mainly adjusted in the underlying Joplin Member. Ore may occur in such places, especially along the partings between the individual chert beds. Where deformation was intense enough to form fractures and breccias crosscutting L bed, these breaks were also commonly mineralized. L bed ore is comparatively richer in galena, which seems to be less dependent on carbonate rock as a precipitant and more dependent on open spaces than sphalerite. Such ore is commonly mined contemporaneously with subjacent M bed ore, or possibly at a later date from M bed workings, but it would not be rich enough for exploitation if it were isolated. The contact of L bed with the underlying stratum is an especially common location for a band of galena.

In relatively pure limestones without interbedded cherts, the favorable chemical character is offset by an unfavorable structural one in that the limestone deforms plastically without fracturing. Such limestones can be mineralized only by diffusion of mineralizing solutions from adjacent fractured ground. The Chester strata exemplify these conditions. Over most of the field they have been sealed off from the ore solutions, but locally, as along the axis of the Miami trough in the Ritz, Crystal, and Central mines, ore solutions have gained access apparently through fractures in the subjacent Boone, and the Chester has been richly mineralized. Quartzitic sandstone beds in the Batesville above a condensed Hindsville section here have undoubtedly also contributed a certain favorable brittleness. As the Chester is comparatively thin here, the total remaining thickness is mineralized over parts of these tracts.

Cotton-rock zones in the upper part of the Boone have in general yielded to stresses much as limestones, and they have not been extensively mineralized, except where fracturing has been intense over a thick vertical interval. Similarly, L bed and the top part (N bed) of the Grand Falls Chert Member, where cotton rock, have lacked the rigidity to transmit stresses capable of breaking the adjacent ore beds, and consequently the adjacent ground is not generally mineralized.

The Moccasin Bend Member of the Boone carries ore chiefly in ground in which K or M bed has been mineralized, and the lower beds of the sequence (G–H beds) are more commonly mineralized than the uppermost beds. This fact suggests that the main feeding channels for the ore solutions were in K and M beds, with variable penetration at different places to the higher levels.

Structural Characteristics of the Ore Deposits

A few ore deposits in the Picher field were obviously localized along the Miami trough, particularly at the southwest side of the field (pl. 1). However, a large part of the district production has come from mines lying southeast of the trough, well beyond the limits of the structural displacement. Although most of the mines are within 2½ miles, the most remote deposits lie nearly 6 miles from the trough. On the northwest side, excluding the isolated minor deposits southwest of Melrose, mineralized ground extends a maximum of about 3 miles from the trough. Some of the ore deposits trend parallel to the trough, and some at a high angle to it, approximately parallel to the Bendelari monocline, but these trends are not pronounced except in the northwestern part of the field. In the rest of the field, however, there is an obscure large-scale pattern of alinement of ore bodies along or near these two trends, regardless of the trends of the individual ore bodies (pl. 1). The pattern suggests relationship to deep-seated linear breaks which become subdued, discontinuous, and partly rotated in other directions at the level of the ore deposits, though maintaining general linearity parallel to the deep breaks.

The Miami trough and Bendelari monocline are zones of recurrent major failure from regional stresses probably in the Precambrian basement rocks. Other parallel, but less pronounced, breaks in the structurally competent rocks at this deep level transmitted a complex pattern of stress into overlying strata. The minor local deformation through the mining field resulting from this stress during the post-Krebs deformation was important in localizing ore deposits because of the brittleness of the rocks involved in the deformation. Any slight vertical warping or differential lateral thrust of structural blocks produced, at different levels in the Mississippian strata, zones of jointing that may extend through several of the lithologic units. Such zones show little displacement and little consistency as to trend or length. They commonly trend parallel to the structure contours drawn at stratigraphic levels near that at which the joints are developed, but there are many exceptions to this generalization. The broken and shattered ground formed by the structural movements along joint zones and less definite loci of rock failure was ideally suited for invasion by the ore solutions. Some of the earlier channels of underground solution movement established in the period of pre-Krebs erosion, where not silted up, were undoubtedly also utilized by the ore solutions in gaining access to ground susceptible to mineralization. The routes by which the ore solutions moved through the ore-bearing strata were devious, and probably were as commonly horizontal as crosscutting at this general level.

An earlier deformation, pre-Krebs in age, affected both the Miami trough and Bendelari monocline, during which the limestone in the chief ore-bearing beds deformed plastically, though the contained chert was extensively brecciated. The ore deposits have structural features relict from both periods of deformation.

The form of a deposit differs somewhat, depending partly on the stratigraphic unit in which it occurs. In most of the ore-bearing strata the ore bodies are elongate and have a curvilinear trend following the joint systems, or following irregular channels of easy solution movement in broken ground. Because the strata are generally fairly flat, the floor of such an ore body, called a run, is typically nearly level, and the roof maintains a fairly constant height, provided the ore stays within definite stratigraphic limits. However, runs are also common on structural slopes, even as steep as the flanks of the Miami trough and Bendelari monocline. The fracturing and shattering in ground immediately bordering faults, such as those along the Miami trough, may be favorable to mineralization, and the floors of the resultant ore runs are commonly inclined, either because the strata are titled or because the ore fails to stay within stratigraphic boundaries. The few runs localized along faults are straight, as on the east forty of the K. E. Jarrett tract in Kansas, but those controlled by less pronounced structural or solutional features may be irregularly curved or even circular around a barren core.

A type of ore run that has been of major productivity in those parts of the field containing gray spar dolomite is localized in broken ground on the peripheries of dolomite cores, especially in M bed. These cores may be nearly circular, as, for example, one on the northern part of the Midcontinent tract (pl. 5), but more commonly they are irregular or elongated. They are generally several hundred feet across, and the more elongate ones may extend for more than a mile. One of the most remarkable linear cores that has been mapped extends—although with imperfection of detail—from the Kansouri tract southwest to the Cherokee, thence west and northwest through the Bendelari,

and north-northwest through the Tulsa-Quapaw, a distance of about 7,500 feet (pls. 5, 6). In this course it is first parallel to the Miami trough, then curves to parallel to the Bendelari monocline, and at its end, curves still farther to an acute angle (about 25°) with the monocline. The ore runs, which are commonly only a few tens of feet wide, follow the borders of the dolomite cores, in places circling completely around them, but elsewhere showing gaps of varying extent. Locally, the mineralized border zone may be broad, transgressing into the dolomite, or the dolomite border may be indefinite. Some of the cores are big enough and irregular enough to contain inclusions of nondolomite, and ore runs may be fairly well defined along the interior borders between the dolomite and these internal inclusions, as, for example, along the west end of the property line between the Cherokee and Chubb (pl. 6).

The ore runs around dolomite cores have a mineralogic asymmetry of their opposite walls (discussed in the next section of this report) that can only be explained on the assumption that the mineralizing solutions gained access to the ore bed at the centers of the dolomite cores. Physical or chemical conditions were not favorable for ore deposition at the centers, but became favorable when the ore solutions had migrated a sufficient distance outward from the centers. Thus, the feeder channels that admitted these particular ore solutions are not the scattered fractures that now show in the ore runs, but are those cutting the centers of the barren dolomite cores where commonly they are not exposed. Longitudinal fractures that show locally in the roofs of the circumferential ore runs may be in large part secondary, an effect rather than cause of the mineralization. They were commonly produced by slumps due to differential volume changes in the blocks of ground along and bordering the ore runs during mineralization.

The dolomite core in places contains enough ore that parts or all of it have been mined out. In other places the core had been so thoroughly leached and compressed to a tight chert mass that not enough dolomite matrix remained to precipitate commercial quantities of ore. Introduced Pennsylvanian shale and residual clay that formed during the leaching added impermeability to the core, great enough locally to have insulated and protected small blocks of unaltered limestone. However, the shale and clay are not confined to the dolomite cores, but they also occur in surrounding areas. Intense leaching in the centers of some dolomite cores has obscured the structural breaks that admitted the ore solutions here, and some of these dolomite cores may be centered on pre-Krebs solution channels that were not related to recognizable fracture patterns.

Where ore occurs in dolomitized ground on both the K and M beds and locally higher levels, as on the Tulsa-Quapaw, Bendelari, and Midcontinent tracts, the gray spar blocks are roughly coextensive on the several levels, though the boundaries do not superpose in detail (pl. 11). In these examples the ore lies along the border of the dolomite on the M bed level, but commonly spreads through the complete width of the dolomite block on the K bed level (pl. 11).

Circular or elliptical ore runs around dolomite cores are also conspicuous in the Joplin, Mo., part of the Tri-State region, and they occur in the Aurora and Granby districts of Missouri (Smith and Siebenthal, 1907, p. 15–16, pls. B, D–F). The deposit at the Zig Zag mine, near Joplin, is concentrated chiefly on a contact between gray spar dolomite and unaltered limestone, and is thus presumably on the border of such a dolomite core.

In beds favorable for ore or in areas widely affected by jointing, ore deposition may spread over greater widths than normal, and the ore runs grade into blanket deposits that have very irregular lateral boundaries. Many of the ore bodies in M and K beds in the central part of the field are large blanketlike masses, some of which are more than 1,000 feet across. Some were mined completely except for small pillars left for support. In some of these bodies, such as in the Netta mine (pl. 8), ore in the central parts contained a dolomite gangue partly replaced by jasperoid, and was surronded by jasperoid-bearing ground that formed at least half of the ore body. In other blanketlike ore bodies, as in the Blue Bird mine (pl. 9), the gangue material was almost entirely jasperoid.

The sheet-ground deposits in the Grand Falls Chert are low-grade blanket deposits that have been of great commercial value because of their wide areal extent. Sheet ground typically has little or no structural relief, and the jointing which presumably furnished access for the ore solutions is weak and inconspicuous. Possibly cracks along the bedding may have contributed as much as vertical jointing to the preparation of the ground for mineralization. The more extensive sheet-ground deposits are at the edges of the field and have no ore deposits in overlying strata. In the southwestern part of the field, the sheet-ground deposits lie adjacent to deposits in the Joplin Member, but there is very little overlap (pl. 9).

In a typical M or K bed deposit, the primary chert has been irregularly brecciated to varying degrees in different parts of the deposit. Commonly, the chert nodules and lenses have been shattered to angular frag-

ments of all conceiveable sizes and shapes, and so rotated within the matrix that no suggestion of their original position remains. The stratification of the bedded chert in the lower part of L bed may likewise be shattered and destroyed, and such chert—contributed to the underlying breccia,—is for the most part indistinguishable from the M bed chert. In exceptional instances thick slabs of lower L bed may be plucked out and are now recognizable where they have been rotated laterally into a position even underneath unbroken basal L. These ore-bed breccias are of tectonic origin, wherein horizontal slippage has been concentrated in the structurally weak ore bed during the relatively slight warping to which the mineralized block has been subjected. The brecciation varies in intensity from place to place, from the described extreme to relatively unbroken ground in which the chert has been cracked but still remains virtually in place. Some of the richest ore is in unbroken ground on the borders of brecciated blocks, where the mineralizing solutions have diffused into the coarse limestone between the chert beds and nodules. No system is recognizable in the pattern of brecciation, perhaps because the degree of brecciation is not readily mappable. The chert and limestone show no details of bedding structure that would allow distinction between thorough brecciation and a comparatively mild shattering in which just enough movement had taken place to destroy the continuity of individual beds or nodules of chert. Or, as the brecciation of the chert took place within a relatively plastic medium, small structural units within this medium could be rotated in a thoroughly brecciated block with very little brecciation of enclosed chert nodules. Thus, chert nodules oriented in all positions, even vertically, and broken in several places but without enough displacement to destroy recognition of their continuity, are of common occurrence.

In the later mineralization the chert was practically unaffected, although fossils enclosed within it were either leached or replaced by jasperoid or, in a few places, by sphalerite. Only rarely was the chert replaced by sparse disseminated sphalerite, and in such instances the band replaced was probably one of the originally more limy bands in the chert. The limestone matrix enclosing the chert, however, was completely replaced by some phase of the mineralization, most abundantly by gray spar or jasperoid, but also by the ore and by certain of the other gangue minerals. In this process more material was leached than was deposited, with the result that vugs and solution caves are abundant, and the mineralized stratum is commonly condensed in thickness.

In those blocks where the host rock was comparatively undisturbed, there was a tendency for different phases of the mineralization to be banded parallel to the bedding. The texture of the gray spar dolomite that has replaced the limestone commonly varies gradationally from bed to bed, or the amount of dark residual clay interstitial to the dolomite grains may so vary. Certain bands were more completely replaced by disseminated sphalerite than others, with gradational boundaries between bands, and beds as much as 2 feet thick were locally wholly replaced. Usually, however, the replacement was incomplete, and consisted both of finely disseminated sphalerite and of irregular replacement pockets of more coarsely crystalline sphalerite or galena, with or without pink spar or calcite, in the gray spar or jasperoid. Pink spar and calcite seams, varying abruptly in thickness, tend to form along the bedding, and the galena likewise tends to segregate in the seams at the tops of the sphalerite bands. Locally, gray spar and jasperoid may be interbanded, though with incomplete segregation of the two (fig. 42). Such banding is, at least in part, inherited from an initial banding of gray spar in limestone, the jasperoid later selectively replacing the remnant zones of limestone. This banding is similar in origin to that observed in greater perfection in the ore deposits of northern Arkansas where, however, the dolomite bands were probably formed in the limestone during the sedimentation process rather than as an early phase of the mineralization (McKnight, 1935, p. 112, 119). Much of the interbanded gray spar and jasperoid also contains thin bands of pink spar, which have grown on the lower borders of the gray spar bands and face euhedrally into the underlying jasperoid. Pink spar is rare at the upper borders of the gray spar bands, though disseminated sphalerite tends to concentrate there.

The thinnest banding is that found locally in massive jasperoid where it is expressed as thin alternating bands of dense and more porous jasperoid (fig. 38). Such banding is doubtless inherited from an obscure and unsuspected primary layering in the original limestone that is accentuated by the jasperoid replacement.

The banding parallel to the bedding, as brought out by the mineralization, is irregular and discontinuous. It is crossed at all angles and at various trends by equally irregular and discontinuous veins and veinlets of pink spar, calcite, or jasperoid, the first two of which may contain irregular blebs, lenses, and pockets of sulfides. The mineral and gangue may completely fill the space between the vein walls, or vugs and water channels may allow the development of well-crystallized material along their walls. Most veins are less than 6 inches thick, but they vary so abruptly in thickness along their course

as to indicate that replacement has been dominant in their formation. Commonly, they do not extend beyond the confines of the ore bed, though some in M bed may break up into the lower part of L bed where they are usually deflected and die out along the bedding. J bed forms an equally effective seal at the top of K bed.

In a few places the initial gray spar replaced the limestone only a short distance from crosscutting fractures along which it was introduced. The remaining limestone was later selectively replaced by dark jasperoid, accompanied by pink spar "reaction rims" at the gray spar contact. The general effect simulates a stalactitic growth of gray spar hanging, however, in solid jasperoid. In figure 42, some of the ground between the gray spar "stalactites" was also partly replaced by gray spar along bedding bands during the initial dolomitic replacement.

In brecciated ground the banding parallel to the bedding does not exist, but the cross veining by various phases of the mineralization, as well as the pockety replacement by sulfides and carbonates, is similar to that in unbroken ground, though perhaps even more irregular.

The fracturing that controlled cross veining in ore beds is a manifestation of the same tectonic deformation that produced more regular jointing in competent units, such as L bed. However, the two types of break are not necessarily coextensive. The fracturing within the ore beds that permitted access of the ore solutions is commonly too irregular to map, or has been obscured or removed by mining operations, and the joint systems in the competent units may be obscure. Joints in L bed are generally tight and unmineralized except for scattered specks of ore or gangue minerals; but locally, thin steeply dipping veins may be present along some of the joints, and in a few places irregular veins relatively rich in lead have been exploitable. The fracturing related to ore was in fairly brittle rock as compared to the physical state of this rock when much of the tectonic brecciation occurred. Either the ore beds were less plastic at the time of the final joint fracturing, or the stresses were applied differently so that the carbonate rock fractured instead of flowed, as it had done during the earlier deformation.

The vugs, which are ubiquitous in the ore deposits, date from the time of mineralization. They show all gradations in size and shape to caves that may be several tens of feet long and several feet high. In dolomitized ground the caves are lined with pink spar, as well as any or all later minerals. In another type of ground, late calcite tends to predominate and to cover up any earlier minerals that may have originally lined the caves. Although many of the smaller drusy openings are obviously along, or adjacent to, cross fractures, the larger caves are evidently controlled by the greater solubility of certain strata. Many vugs and small caves occur in massive dolomite or jasperoid with no obvious control other than the intense corrosion to which all the mineralized ground has been subjected. Localized solution of the originally calcareous material throughout all or part of M bed commonly condensed the original thickness and allowed the insoluble chert to settle away from the structurally competent L bed roof; hence the commonest locus for caves is at the top of M bed.

The outer sides of the ore runs surrounding dolomite cores are, in general, more open and vuggy than the dolomite sides, owing to excess solution of the original carbonate rock over replacement by gangue and ore minerals. Farther out on their fringes, these deposits commonly grade into, and are bordered by, a characteristic type of barren ground formed by complete or nearly complete solution of all originally calcareous material in the ore bed, which leaves the residual chert loosely piled together in the so-called boulder ground or boulder piles. Such boulder ground is generally of wide extent, and the superincumbent beds, lacking support, have slumped down evenly onto the boulder ground without formation of true caves. Where this ground is formed in unbrecciated terrain, the original chert beds, lenses, and nodules remain intact and preserve an especially coarse porosity.

In beds that contain a higher proportion of chert, such as the Grand Falls sheet-ground zone or the G-H bed zone at the base of the Moccasin Bend Member, the ore structures are somewhat simpler than in M and K beds. Brecciation is not nearly so common, and where present, is likely to be confined to local cracking or shattering of chert beds or lenses, without the profound brecciation that in places affects the beds containing a relatively larger percentage of limestone.

The fracturing that opened the ground to the mineralizing solutions was largely spent by the time the mineralization was effected. However, local minor fracturing and brecciation persisted into the warning stages of mineral deposition. Mild, but widespread, fracturing at a relatively late stage commonly broke across the earlier pink spar, gray spar, and jasperoid that contain disseminated sphalerite and galena. The free broken faces of these fractures exhibit only the small insignificant crystals of sphalerite, galena, chalcopyrite, marcasite, or pyrite that were deposited during the waning stages of the sulfide mineralization. However, such fractures may be extensively filled by calcite. The minerals in these late breaks commonly have precipitated selec-

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tively on certain materials that were exposed in the cross fractures; for example, chalcopyrite and marcasite may appear only on the broken pink spar, or fine sphalerite only on the disseminated earlier sphalerite. The leaching of galena (p. 125) is commonly associated with late fracturing, and quite possibly was facilitated by such fracturing, which exposed to later solutions the galena precipitated at an earlier stage and which was unstable in the later environment.

Zoning of Ore Deposits

Although the deposits of the Picher field are not zoned on the scale recognized in western mining districts where successive zones are hundreds or thousands of feet apart, parts of the field that contain gray and pink spar dolomite exhibit sharply defined mineralogic zoning around the gray spar cores. The richest concentration of ore is on the periphery of the dolomite mass, so that, at least in the early stages of mining when only the richest ore was taken, the mine working is a curving linear gallery that, within the ore bed, makes a complete or partial circuit of the gray spar block. Plates 5–10 show the relation of stoped ground to the gray spar border over a large part of the mining field.

A pronounced asymmetry exists between the two sides of a run bordering a gray spar mass. On the inner side, gray and pink spar dolomite are the prevailing gangue minerals, though the gray spar may be partly replaced by jasperoid. The richest sphalerite and all the chalcopyrite, enargite and luzonite are concentrated on this side. All types of sphalerite are present, but the black jack variety is virtually confined to this dolomite side. Galena, marcasite, and calcite, though present on the inner side, are in relatively greater concentration on the outer side, where jasperoid and calcite are the gangue materials. Sphalerite is still present on the outer side, but relatively less abundant; at the outer edge of commercial ore it is characteristically in small pinhead-sized crystals of rosin or ruby jack. The original chert of the ore bed is a constant constituent of both sides. The ore-bearing part of the zonal sequence is commonly encompassed within a distance of 100 feet or less. The more open and vuggy outer side of the ore run commonly grades farther outward into barren boulder ground, characterized by fragments or slabs of loose residual chert from which all matrix material has been leached, except for thin rims or films of jasperoid. This boulder ground contains some marcasite, calcite, and scattered small crystals of galena and sphalerite but in less than commercial concentration. Boulder ground is puzzling in that it represents a gross porosity that was available for ore deposition, yet it contains no ore except along one border where it approaches the ore run. Evidently, factors other than the availability of space determined whether or not ore was deposited. Beyond the boulder ground is the unaltered limestone and chert that have been unaffected by the mineralizing solutions.

Except for discrepancies in the copper-bearing minerals, the mineralogic zoning around the gray spar cores roughly duplicates in space, from the core outward, the paragenetic order of crystallization. The copper minerals are coextensive with the dolomite in space, but chalcopyrite, and particularly enargite and luzonite, are later than, and well separated from, the dolomite in time of crystallization. Jasperoid occupies a wider territory in the zonal relationship than is shown on the paragenetic chart (fig. 41); but with dolomite, sphalerite, galena, marcasite, and calcite, the zonal and paragenetic sequences are the same. It is as though a nearly constant differential in the physical or chemical environment in the locus of deposition was maintained during the main surge of mineralization, which determined the positions of main mass deposition for each ore and gangue mineral; whereas in the later stages of waning mineralization the zones in the depositional environment migrated back toward the dispersion center, so that remote minerals in the zonal sequence were deposited on the inner ones as later minerals in the paragenetic sequence.

In the discussion of paragenesis (p. 129–130) it was pointed out that sphalerite was mostly deposited earlier than galena, but there are numerous reversals in this order of crystallization involving small late crystals of sphalerite which, quantitywise, account for a very minor percentage of the total sphalerite. Although these reversals may be found on the dolomite side of ore runs bordering dolomite cores, they are much more characteristic of the distal jasperoid-galena margins of the ore where the only sphalerite present is likely to be in very small rosin or ruby jack crystals.

Origin of Deposits

Any discussion of the origin of the Picher ores must recognize the similarity in mineralogy and mode of occurrence throughout the Tri-State region, including also the Central Missouri district lying southwest of Jefferson City, Mo. This similarity implies a common origin. As pointed out by Ohle (1959), there is a worldwide group of ore deposits of the Mississippi Valley type which possess many features in common and which probably have a similar mode of origin. However, within a framework of generally mild structural deformation, carbonate host rock, simple mineralogy, and absence of igneous evidence, variations occur between

individual districts, such as the relative proportions of the different metals contained and the depositional environment of the sedimentary host rocks. Thus, the southeast Missouri lead region differs from the Tri-State region in the relative proportions of the metals, the presence of appreciable nickel and cobalt, and the existence of buried Precambrian knobs in the depositional environment. The discussion of origin that follows will refer specifically to the Tri-State and Central Missouri deposits, but will be consistent with the broader field so well defined by Ohle (1959).

The literature on the origin of the Tri-State ore deposits is extensive. In 1939 a review volume entitled "Contributions to a knowledge of the lead and zinc deposits of the Mississippi Valley region" was published by a subcommittee of the Committee on Processes of Ore Deposition, National Research Council, under the editorship of Edson S. Bastin (Bastin and others, 1939). This volume critically reviewed the different theories of origin that had been advocated, though it did not review historically the development of these theories. Ohle (1959) made a similar critical review, but from the broader viewpoint of worldwide occurrence of the Mississippi Valley type of deposit.

Ohle discussed five modes of origin that have been advanced to explain the Mississippi Valley type of ore deposits, though not all have been applied specifically to the Tri-State ores:

1. Original syngentic deposition.
2. Original scattered deposition of the metals with later concentration by regional metamorphism.
3. Original scattered deposition of the metals in sedimentary rocks with later concentration by circulating meteoric water moving down.
4. Original scattered deposition of the metals in sedimentary rocks with later concentration by circulating meteoric water moving up (the artesian circulation theory).
5. Deposition from fluids of igneous derivation.

An additional mode of origin, called the connate-hydrothermal, has recently been defined by Hall and Friedman (1963) to explain the Upper Mississippi Valley zinc-lead deposits and certain temporal stages of the ore deposits in the Cave-in-Rock district, southern Illinois. All these deposits are considered by them to be of the Mississippi Valley type. The theory of origin is an elaboration of White's (1958) postulate that the Tri-State ores were deposited by connate brines of the type found in oil fields. The hypothesis of origin in its extreme form is as follows: Original scattered deposition of the metals in sedimentary rocks with later concentration and deposition from warm upward-moving connate waters, heated either by deep burial or by proximity to a buried igneous intrusive.

Many features in the geology of the Tri-State deposits are not consistent with the first four and the connate-hydrothermal modes of origin, whereas there is considerable evidence, much of its circumstantial, favoring an igneous derivation, which is the origin advocated in the present report.

SYGENETIC, METAMORPHIC, AND METEORIC HYPOTHESES

As objections to mode 1 in the preceding list, Ohle (1959) has pointed out the occurrence of the ores in crosscutting veins or in beds of nearly solid sulfide and their relation to faults and other secondary structural features, all of which are inconsistent with primary sedimentation. To these objections could be added the abundance of fossils in the ore-bearing strata, indicating a healthy undisturbed environment for organisms.

As objection to mode 2, he cites the lack of any regional metamorphism in the Ozarks that could account for the ore concentration in post-Pennsylvania time. Not considered by him is the low-grade regional metamorphism produced in the Ouachita Mountain geosyncline by the orogeny that culminated near the middle or end (p. 82) of the Pennsylvanian Period (Goldstein and Reno, 1952; Flawn and others, 1961, p. 121–124). Whereas the Picher ore deposits are of post-Krebs age (p. 154), much of the Ouachita orogeny was pre-Krebs (King, in Flawn and others, 1961, p. 186–188); hence only the metamorphism during the culminating stages of the orogeny could be considered. Although this metamorphism was undoubtedly adequate to produce thermal solutions, the difficulties in concentrating the metals in these solutions and transporting them 150 miles laterally to the Picher field are greater than in calling upon some more local but deep-seated crustal source.

The theories of concentration by ground water (modes 3 and 4 of the list), as applied to Tri-State ores, postulate the original source of the metals as the Precambrian igneous rocks, from which they were weathered and dispersed, in trace concentrations, into the Pennsylvanian shales according to one theory (mode 3), or into the Cambrian and Ordovician dolomites, according to the other theory (mode 4); subsequently the metals were leached from these sedimentary rocks and concentrated into ore deposits by ground waters during a second erosion cycle. Ohle (1959) and Bastin and Behre (Bastin and others, 1939) have pointed out difficulties in these theories. Questions raised by them that are applicable to Tri-State ores

include the lack of evidence as to an adequate source of the metals in the Precambrian rocks, particularly in view of the dissipating action of the first erosion cycle; the inefficiency of ground water in leaching the source rock during the ore concentration cycle, with resultant inadequate supply of metal and necessity for high efficiency in precipitation; the insolubility of lead in any conceivable leaching solutions; the high ratio of readily leachable copper to lead in the postulated source compared to its negligible proportion relative to lead in the ore deposits; the inadequacy of the proposed ore-precipitating mechanisms; and the lack of any mineralogic resemblance of the ores to the usual products of secondary enrichment by meteoric waters. Despite the difficulties raised, Ohle cautions against premature rejection of these lateral secretion theories, though he stresses the greater flexibility of the magmatic-hydrothermal theory (mode 5).

The theory of ore deposition by circulating artesian waters (mode 4), as developed by Siebenthal (1915) specifically for the ores of the Tri-State district, has met with varying acceptance. It presumes that, at a late stage in the erosion of the Pennsylvanian shale from the Ozark dome, the metals were dissolved from disseminated minerals in the gently dipping Cambrian and Ordovician strata by waters that entered the outcrops of these strata high on the dome and were deposited in and near the top of the Mississippian limestone on the borders of the dome by the rise of artesian waters to the zone of escape at the inner edge of the shale outcrop. The cause of precipitation was believed to have been the loss, at the zone of artesian escape, of the carbon dioxide which held the metals and gangue minerals in solution as bicarbonates in the presence of hydrogen sulfide. Figure 44 shows the presumed circulation of ground water at the time of ore deposition. Artesian water, beginning at points A, B, and C, on the outcrops of strata below the shale, follows the courses of the dashed lines, rising to the surface at D, where, upon escape of carbon dioxide, the mineral matter in solution is precipitated. Especial emphasis is placed on the belief that the ores lie comparatively near the surface and near the inner edge of the overlapping Pennsylvanian shale.

Siebenthal's theory was published at about the time that the greatest strides were being made in extending the boundaries of the Picher field northwestward beneath the edge of the shale. Many rich new deposits were eventually found several miles back from the main edge of the shale, though there is a window through the shale along and east of Tar Creek (pl. 2) which brings an edge considerably closer to some of the ore deposits. Nevertheless, numerous mines below shale cover in the northwestern part of the field lie 4 miles from the nearest known shale edge, and the farthest deposits on the Karcher tract are more than 5 miles back from the edge. This seems a prohibitive spatial separation of cause from effect in the proposed hypothesis.. Furthermore, the farthest remote deposits are 2–3 miles beyond the Miami trough, and as pointed out by Scoff (Reed and others, 1955, p. 34), the Miami trough is believed to be a hydrologic barrier to the movement of ground water in the Ordovician rocks that are most involved in the artesian theory..

The deposits at Picher are not centered on the Tar Creek window as might be expected if deposition were from artesian solutions moving upward through this window (fig. 45). Nor do they occur at the surface in the Chester limestones that crop out in this window, where the greatest concentration should be expected if the precipitating mechanism were the escape of carbon dioxide at the surface; instead, they occur precisely as in other parts of the field, most in the Joplin Member but some in K bed, and 100–240 feet below the surface at different places, depending on the structure of the ore-bearing beds. There is a similar lack of correlation between the richer ore deposits and the main edge of the shale at the east side of the mining field and a similar lack of ore in the Chester limestones along the shale edge.

Although the deposits in the Melrose area are more than 7 miles from the window on Tar Creek, some of them lie only 3 miles from another window of unknown size (because largely concealed by Neosho River alluvium) on Fourmile Creek 1 mile above its mouth (pl. 2). As at Picher, the deposits show no apparent genetic relationship to this window.

Studies of liquid inclusions trapped in the ore minerals of the Tri-State region show that the depositing solutions were concentrated brines of sodium and calcium chlorides (Newhouse, 1932; Roedder, 1963). The concentration of these brines is much greater than that of sea water and is incompatible with derivation of the depositing solutions from meteoric waters.

CONNATE-HYDROTHERMAL HYPOTHESIS

The composition and concentration of the fluid inclusions in the sphalerite and galena of the Tri-State region are similar to those of certain deep oil-field brines, which led White (1958) to postulate that the ore-depositing solutions were, indeed, mobilized connate solutions from some deep sedimentary source. However, sodium chloride, the chief dissolved constituent of the fluid inclusions, is also a major constituent of magmatic solutions (White, 1957a, p. 1643, 1644–1647)

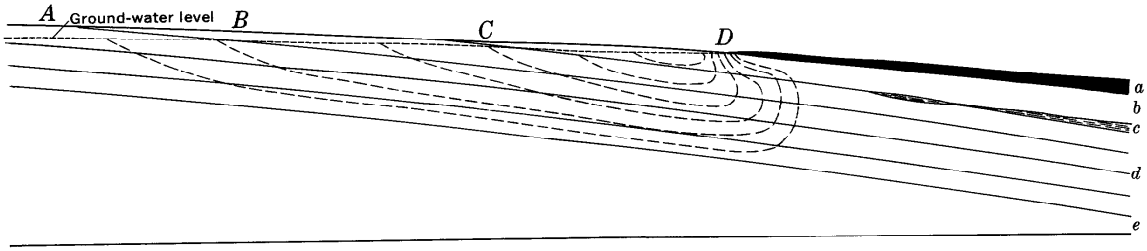

FIGURE 44.—Diagram showing a late stage in Ozark artesian circulation. *a*, Pennsylvanian shale; *b*, Mississippian limestone; *c*, Chattanooga Shale (Devonian and Mississippian); *d*, Ordovician and Cambrian dolomites; *e*, Precambrian rocks. After Siebenthal (1915).

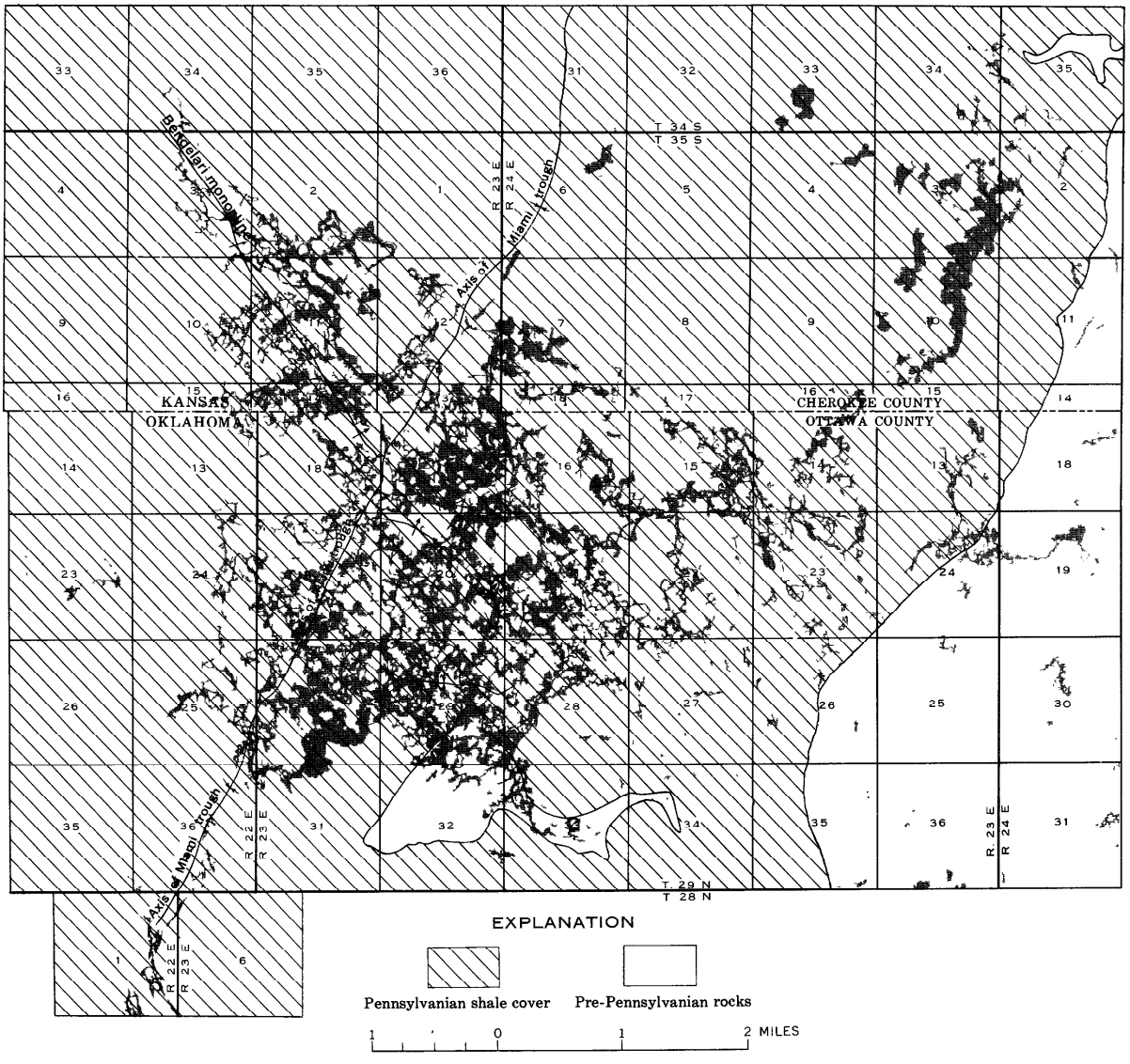

FIGURE 45.—Mine workings in the Picher field in relation to the Pennsylvania shale cover.

and of volcanic sublimates (Fenner, 1933, p. 92–95). Roedder and others (1963, table 4, p. 368) included sphalerite from unquestioned magmatic sources as well as from the Picher field in their study of fluid inclusion compositions. The calcium-sodium ratios are comparable in the two types of deposits, though the total salt concentrations average somewhat higher in the fluid inclusions from Tri-State sphalerite (see also Roedder, 1963, p. 177–178).

The liquid inclusions are only a partial record of fluid composition during the sequence of mineral deposition, if the gangue minerals as well as the ore minerals are considered. The period of mineralization in the Picher field began with deposition of calcium-magnesium carbonate, and was followed by deposition of silica in the form of jasperoid, which preceded most of the sulfides (fig. 41). In actual mass, silica was the major constituent carried and precipitated during the mineralization cycle. Silica as vein quartz is a characteristic gangue in magmatic ore deposits that were formed at considerably higher temperatures than the Picher ores, and magmatic solutions undoubtedly retain their saturation in silica into lower temperatures. Thus, hot springs of volcanic origin are commonly still saturated with silica at the relatively low temperatures prevailing at the vents (White, 1957b, p. 1673). Oil-field brines, on the other hand, have "nearly the lowest SiO_2 content of all natural waters" (White, 1958, p. 1661). Although the period of silica deposition in the Picher field largely preceded that of the ore minerals, it seems to us incongruous to presume that the original connate solutions were flushed out of the host rock by carbonate- and silica-bearing solutions which deposited some of the gangues, and that connate brines were later reintroduced to deposit the ore minerals.

The resemblance of the fluid inclusions from Tri-State sulfides to oil-field brines is believed to be superficial and of no genetic significance. The concentrations of the major components in both instances are doubtless controlled by saturation solubilities at some stage in the complex histories of the respective solutions.

MAGMATIC-HYDROTHERMAL HYPOTHESIS

Several independent evidences suggest a hypogene source for the bulk of the ore solutions. These include the existence of a definite order of mineral crystallization with parallel zoning, comparable to that found in areas of unquestioned magmatic-hydrothermal deposits; an indicated single episodic surge in mineralization, as suggested by certain features in the paragenesis; an indicated considerably higher temperature of the solutions than could be acquired from any surmised deep ground-water circulation; an indicated high brine concentration in the ore solutions as determined from a study of fluid inclusions; and the similarity in composition of the ores to other ore deposits whose genetic connection with intrusive igneous rocks is generally recognized. All the evidences considered except the last are indicative only of hydrothermal solutions, but as pointed out by Ohle (1959), these are not necessarily synonymous with hypogene solutions. However, the difficulties of explaining hydrothermal solutions without recourse to a hypogene source are so great as to make the hypogene origin the most plausible hypothesis.

ORDER OF CRYSTALLIZATION AND ZONING

In descriptions of ore deposits, a definite order in the crystallization of the minerals is commonly pointed out for those that are genetically related to cooling igneous intrusives, and this order is roughly the same in different deposits. Furthermore, many mining camps show a mineralogic zoning in space around instrusives that is roughly the same, from the intrusives outward, as the order of crystallization (Park, 1955, p. 233–34). Commonly, the mineral composition and sequence in the outer zones of such camps are comparable to those in the Picher mining field. The zoning in space is most easily explained by differing solubilities under a temperature gradient; and the orderly change, with time, in the kind of mineral being deposited at a given place is most easily explained as a result of changes in solubility with falling temperature, which is accompanied by migration of the temperature zones and resultant mineralogic zones back toward the source of heat. Should the changes in minerals being precipitated at a given point be more directly attributable to changes in the composition of the solutions with time, these changes are most probably related to cooling nearer the ultimate source. The interrelated phenomena—zoning and paragenetic order—are the logical accompaniment of cooling hydrothermal solutions that owe their heat to derivation from an igneous source, and are hard to explain under any other theory.

Spatial zoning in the Tri-State ores is not around any igneous mass, but around dolomitic cores that doubtless mark the main channels of solution movement. Thus, the heat source as it affected the wallrocks during the time of mineral deposition was local and sharply defined; and the temperature gradient set up in the rocks was much steeper, with resultant condensation of mineralogic zoning into a much shorter space, than where an intrusive igneous body is the immediate center of zoning. The centers of the dolomite cores were too hot to precipitate the ore minerals when most of

the ore solutions were traversing the ground, and any migration of the ore-mineral zones back into the dolomite centers in the dying stages of mineralization usually produced ore of too low a grade to be workable. In those places where the dolomite core is mineralized throughout its width on the K bed level above a barren core center on the M bed level, the zone of ore precipitation is in the shape of a flat-topped shell, when considered in cross section.

The temperatures of mineral crystallization in the Picher field is discussed under "Temperature of ore formation" (p. 146), but it is pertinent to mention here the experimental determinations of Schmidt (1962) on the relative temperatures at which sphalerite, pink spar dolomite, and calcite were deposited in the Picher field, for these have a bearing on the paragenetic and zonal sequence. Schmidt found, from a study of bubbles in liquid inclusions, that the pink spar in vugs had crystallized at a lower temperature than the sphalerite. The sphalerite is considered by him to be of two generations: an earlier finely crystalline to massive type that is contemporaneous with the associated gray dolomite and jasperoid, and a later crystalline type deposited in open spaces. The pink dolomite in the open spaces is considered to be of intermediate age, between that of the two sphalerite generations. Schmidt does not give the evidence as to why the gray dolomite and associated [replacement] sphalerite are believed to be earlier than the pink dolomite and euhedral sphalerite in the openings. There is admittedly no evidence, however, proving the contrary, namely, that there is only one generation of dolomite and of sphalerite, other than what seems a logical assumption to us that the minerals replacing the wallrock of a vein are virtually contemporaneous with the same minerals in the vein. Schmidt postulates that the earlier generation of sphalerite was deposited at the highest temperatures (as high as 120° C), the pink dolomite was deposited at lower temperatures, and the second surge of sphalerite in the openings was deposited on the pink dolomite at high temperatures again, though not so high as in the first generation. Calcite was later than the other two minerals investigated and was deposited at the lowest temperatures determined.

This interpretation is more complex than the simple picture of zonal and paragenetic sequences related to cooling of a mineral-depositing system under a diminishing thermal gradient from the dolomite core outward. Furthermore, one set of four sphalerite samples taken across the outer fringe of a dolomite core in the Netta White mine is reported by Schmidt (1962) to show higher temperatures on the outer jasperoid side than on the dolomite side. One cross section is perhaps not enough to establish a reverse thermal gradient from the dolomite core outward as a general phenomenon. However, the temperature relation indicated between the pink dolomite and the later sphalerite in openings, if true temperatures of crystallization are measured, is an anomaly that appears explainable only on the assumption of rising temperature for this stage of mineralization, contrary to the usual implication of paragenetic relations. Deposition of later calcite at the lower temperatures determined by Schmidt is in orthodox accordance with its position in the zonal and paragenetic sequence.

Perhaps the apparent anomaly of higher temperature sphalerite emplaced on lower temperature pink spar in the vugs is explainable if the total thermal history of the dolomite cores is considered. At the beginning of mineralization, the host rock was cold. Upon introduction of hydrothermal solutions through certain centers, the host rock was warmed up gradually as more and more of the hot solutions spread out from the epicenters, and thermal zones were set up concentric to these centers. Calcium carbonate is more soluble in simple water solutions at lower than at higher temperatures, but it is probable that in concentrated brines, such as those depositing the Tri-State ores (see p. 142), relations may be reversed (Roedder, 1963, p. 175); and by analogy, dolomite may reasonably be expected to behave similarly. Thus, dolomite may have precipitated early in some of the inner zones, but as the temperature rose above a certain threshold of tolerance, it was redissolved and moved farther out. The dolomite may have been moved several times before reaching its final resting place. At the climax in the thermal history when the highest temperatures were reached, it is possible that the existing gray spar and pink spar may have been largely deposited at somewhat lower temperatures but were stable enough that they were not readily dissolved. The situation may, however, account for the local partial leaching of the pink spar described on page 128. Similar adjustments to rising temperature may also explain the more severe leaching of some of the sulfides. These must have been more sensitive to the imbalance in equilibrium brought on by changing temperature, for some of the sphalerite, at least, to judge from Schmidt's studies (1962), was crystallized at the highest temperatures recorded by his study.

The preceding explanation is offered to reconcile Schmidt's temperature studies with our independently derived interpretations. The sphalerite could be virtually all of one generation, though that deposited with some of the jasperoid, which preceded sphlarite slightly

in the paragenctic sequence, may have crystallized at slightly higher temperatures than the main surge of sphalerite in the more open vugs. This explanation implies that the crystal structure of the pink spar was sturdy enough to resist the excess pressure created in its fluid inclusions by heating them at least 25°–44° C. beyond their temperature of imprisonment, but an assumption that is only slightly less extreme is implicit in Schmidt's published results.

Single Surge of Mineralization

In discussing the significance of the paragenetic relations (p. 133), the single surge of major precipitation for each mineral was emphasized. This feature is represented diagrammatically by the width of the lenses in figure 41. Any straggling of precipitation beyond the main period was as insignificant crystals that characterized the dying phases of the mineralization, and minor fracturing was sharply bracketed in time within these dying phases. A single brief episodic surge of mineralization is to be expected if the ore solutions are of magmatic derivation, but such a surge is not consistent with the theories heretofore proposed involving deposition from meteoric waters. However, theories of origin from some byproduct of regional metamorphism or from mobilized connate solutions would not be ruled out by this consideration, though they are unlikely from other considerations (p. 141, 142–144).

Temperature of Ore Formation

The temperature at which the ores were deposited has been investigated by Newhouse (1933), who heated thick sections of sphalerite on the microscope stage and noted the temperature at which the gas bubbles in microscopic liquid inclusions disappeared. On the reasonable assumption that the gas bubble represents the vapor phase that appeared in the space formed through contraction of the trapped liquid during cooling, heating would expand the liquid and the bubble would disappear when the volume and temperature at the time of original entrapment were recreated. That the interpretation is sound is shown by the simultaneous disappearance of the bubbles in many adjacent inclusions. Crystallization temperatures of 135°–90° C were indicated by this method for several sphalerites from scattered Tri-State localities.

More recently, Schmidt (1962) has used the same general method to investigate temperatures of crystallization in sphalerite, pink spar dolomite, and calcite from 10 mines in the Picher field. He found that the sphalerite, which he separated into two generations, crystallized at 120°–83° C, pink spar at 96°–76° C, and calcite at 68°–52° C. The highest temperatures for sphalerite (120°–85° C) were found in the "finely crystalline to massive" sulfide of the earlier generation which he considers to be contemporaneous with the jasperoid and gray spar dolomite. In this early sphalerite, differences in the temperature of crystallization of as much as 16° are recorded by fluid inclusions that are only one-twentieth inch apart, interpreted as being, respectively, near the "base" and near the crystal faces of individual [disseminated?] crystals (Schmidt, 1962, fig. 4). The criterion on which the base of such a crystal is determined is not explained. Temperatures of crystallization for the second generation of sphalerite, which consists of euhedral crystals on top of pink spar in open spaces, were 105°–83° C, thus overlapping the temperatures of the first generation but averaging somewhat lower.

Although it is perhaps not necessary to postulate two separate surges of sphalerite mineralization (p. 145), the surprisingly low temperature range found for crystallization of the pink dolomite, compared to that of the sphalerite, makes necessary the assumption of mineral deposition over a span of rising temperature, when paragenetic relations of the dolomite and sphalerite are considered.

The highest crystalization temperatures reported by Schmidt for the "early generation" sphalerite were found in samples from mines nearest the Miami trough, and there was gradation through intermediate mines to the lowest crystallization temperatures found in samples from mines farthest from the trough. A similar temperature gradient relative to the trough was indicated by three more closely spaced samples taken near the trough in the Blue Goose mine. A general decline in temperature with increased distance from the trough during this mineral stage is suggested by Schmidt, though he qualifies this by stating that "the number of samples studied is too small and the lateral range too narrow to justify broad conclusions." These temperature investigations at least suggest that the Miami trough was one major center for dispersal of the hydrothermal mineralizing solutions, and that it evidently offered a rather easy and direct passageway from depth to the site of the ore deposits along it.

The coarser texture and "porphyroblastic" habit of the silica (jasperoid) that is deposited with the ores contrasts markedly with the associated cold water silica (chert), which is much finer grained (cryptocrystalline) and lacks any semblance of axial elongation in the interlocking anhedral grains. The differences in texture are such as might be expected if the jasperoid were deposited at a higher temperature.

Although the temperatures indicated by Newhouse and Schmidt are lower than generally assumed for most

magmatic ore solutions, they are considerably higher than could have been attained at the probable depth of ore deposition under a normal geothermal gradient from the surface. A comparison of the Pennsylvanian rocks in the basins bordering the Ozark uplift (Pierce and Courtier, 1938; Hinds and Greene, 1915) with the meager remnants found on top of the upland (Hinds and Greene, 1915, p. 210) strongly suggests that not more than 750 feet of Pennsylvanian strata (Krebs, Cabaniss, and Marmaton Groups, comprising the Des Moines Series) were ever deposited in the mineralized region, and much of this probably had been eroded by the time of mineralization. Higher Pennsylvanian marine strata in which limestone is conspicuous (Missourian and Virgilian Series) are present in Oklahoma, Kansas, and Missouri many miles out from the Ozark uplift, amounting to a thickness of 1,500 feet in southeast Kansas; but these strata are not believed to have been deposited in the Ozark region, though the evidence is not conclusive. Hinds and Greene (1915, p. 210) state,

> There is still much doubt as to whether the Pennsylvanian sea finally covered practically all of southern Missouri and submerged the Ozarks, though the evidence in hand seems to indicate that a large part of the region was inundated for a comparatively short interval, beginning, probably, near the end of the Cherokee epoch [=end of Cabaniss]. * * * The thinness of the probable marine Pennsylvanian sediments in all of the Ozarks, however, indicates that the sea may have retreated again in a comparatively short time, probably before the end of the Des Moines epoch.

Bridge (1930, p. 149) states that "after the deposition of the early [Middle?] Pennsylvanian sediments the Ozark dome was again warped upward and, so far as is known, never has been resubmerged."

Even if earlier geologic interpretations as to the maximum Pennsylvanian cover over the Tri-State region have been incorrect, a maximum of 2,250 feet, which is the total initial thickness of Pennsylvanian strata in adjacent basin areas, would seem to be the greatest thickness possible. The final uplift of the Ozark area was probably coincident with the Ouachita-Arbuckle orogeny, which was in late Pennsylvanian time (p. 82). There were no deposits of Permian or Mesozoic strata in the area (Eardley, 1962, pls. 8–12). Acquisition of heat by the ore solutions through deep penetration of meteoric waters into the Cambrian and Ordovician dolomites, which total about 1,325 feet, would be unlikely because such a circulation would be contingent on exposure of the dolomites to erosion at a certain altitude on top of the Ozark dome, coupled with a topographic surface at lower altitude in the downdip areas of ore deposition. With an average geothermal gradient of 1°C per 100 feet of depth, 2,250 feet plus the small additional depth that the ores lie below the Pennsylvanian is inadequate to explain the temperatures indicated by the fluid-inclusion studies. The excess heat of the solutions is most easily explained as residual magmatic heat.

Independent evidence that ore deposition took place under shallow cover is found in the extensive flat caves which the ore-bearing solutions leached out of the mineralized limestones and only partly refilled with ore and gangue minerals. Such features are not typical of ground having any considerable rock pressures.

Composition of Ore Solutions

Partial information on the composition of the ore solutions is, of course, given by the minerals precipitated. The sulfides consist of dominant sphalerite, galena, and marcasite and subordinate chalcopyrite, pyrite, and enargite. There is a little wurtzite at Joplin. The gangue minerals consist of jasperoid (which is a form of quartz), dolomite, calcite, and subordinate barite. Translated into the simplest forms as carried in the ore solutions, this means the cations Zn, Pb, Fe, Cu, As, Mg, Ca, and Ba and the anions S, Co_3, and SO_4. However, because of the phenomenon of complexing (Helgeson, 1964), these constituents may be aggregated into many complex combinations of charged and uncharged particles whose relative proportions in the solutions are governed by equilibrium laws dependent on such variables as temperature and relative concentrations of the components. Complexing can account for much greater solubility of the constituents than can the consideration of simple sulfide, carbonate, or sulfate solubilities. The silica was carried in true solution as the little-ionized orthosilicic acid molecule, H_4SiO_4 (Krauskopf, 1956, p. 15). There are traces of other elements, the most interesting of which is nickel, which makes up 0.18 percent of one of the pyrite samples.

The proportions in which these constituents were deposited in the ores bear at most only a crude relationship to the composition of the solutions; for whether a constituent is precipitated or not depends on the solubilities of the possible ionic combinations and not on the concentrations as such. Furthermore, the composition of the solutions may have varied at different times, at least at a given locality of deposition, as witness the paragenetic relationships. For a complete picture of the ore solutions, one would need to know the disposal and composition of the spent solutions. With the deposits forming so close to the topographic surface, as seems most likely, it is conceivable that such solutions may have reached the surface and have been added to the surface drainage.

That there were constituents in the ore solutions which were not precipitated in the ores is shown by Buerger's (1932), Newhouse's (1932), and Roedder and others' (1963) studies of liquid inclusions in the sphalerite and galena. These inclusions contain 10–20 percent of sodium chloride, with lesser amounts of calcium chloride, and a little potassium and magnesium chloride. The inclusions were trapped in the growing crystals and thus are an indication of the composition of the ore solutions. The constituents are the same as prevail in certain "connate" ground waters in areas bordering the Ozark uplift (Siebenthal, 1915, p. 155–157), but the concentrations are many times greater. It seems significant that in both composition and concentration, these inclusions are very similar to those trapped in galena crystals from such typically magmatic-hydrothermal districts as Leadville, Colo., and Freiburg, Saxony (Newhouse, 1932). Fluid inclusions saturated with sodium or potassium chloride, or both, have been reported from several other hydrothermal deposits; those in the quartz from sulfide-bearing veins of the Bingham district, Utah, contain cubes of halite which indicate a concentration of more than 40 percent sodium chloride, in addition to other unidentified constituents, in the solution at the time the inclusions were sealed off (Roedder, 1963, p. 177–178). In some other deposits of unquestioned magmatic derivation, however, the liquid inclusions in sphalerite are more dilute than in those from the Tri-State region (Roedder, 1963). Chlorides of the alkali metals, and particularly of sodium, are abundant constituents of magmas, as evidenced by the sublimates in volcanic regions and by the composition of hot springs in regions of expiring igneous activity (Fenner, 1933, p. 92–95; White and others, 1963). The ratio of potassium to sodium in fluid inclusions from a Picher sphalerite sample is lower than for most typical magmatic waters (compare Roedder and others, 1963, table 4, p. 368, with White, 1960, p. 452), but is comparable to that in fluid inclusions from sphalerite of unquestioned magmatic origin from the Cartagena district, Spain (Roedder and others, 1963, table 4, p. 368).

An additional constituent probably present in the spent solutions was the hydrogen sulfide molecule. When the Picher district was first opened, H_2S was very conspicuous in many of the workings and proved to be very troublesome in development of the mines because of its health hazard. The gas also permeates the domestic water systems of such towns as Columbus, Kans., 20 miles northwest of Joplin, which receives its water from artesian sources in the Ordovician rocks but below an ultimate capping of Pennsylvanian shale. As hydrogen sulfide is a common constituent of ground waters and can be formed in many ways, it would be unsafe to ascribe the wide occurrence of the gas on the Ozark fringe to ore solutions entirely, but the later the ore deposition took place, the more probable that some of this gas encountered in the mines was originally part of the ore solutions.

Mine water that was pumped from the Garrett shaft (NW¼NW¼ sec. 36, T. 34 S., R. 23 E., Cherokee County, Kans., 3 miles north of Treece) in the early stages of dewatering (Feb. 1, 1939) reeked of hydrogen sulfide and contained 0.2 ppm of AsO_4, as determined by K. J. Murata. The water level at this shaft had not been affected by dewatering in the closest mines, 1½ miles to the south; hence this water may be comparable to that saturating the ore at the start of mining in the Picher field.

The high concentration of sodium and calcium chlorides in the ore solutions, in combination with an indicated temperature, accounts for one of the most puzzling features of the Tri-State ores, namely, the primary corrosion of the galena in many places. Sodium chloride solutions are effective solvents of lead sulfide at higher temperatures (Newhouse, 1932, p. 434), chiefly through the formation of chloride complexes (Helgeson, 1964, p. 60–71). With the demonstration that the Tri-State ore solutions were strong brines, the leaching can be explained through either increase in temperature or concentration of the brine at a given elevated temperature, or both. Such changes in equilibrium conditions are to be expected in magmatic solutions.

Under certain conditions, cooling hydrothermal solutions rich in sodium chloride may leach earlier deposited galena without involving a resurgence in temperature or brine concentrations. Helgeson (1964, p. 60–71, 78–79), in a study of the thermodynamic properties of such solutions, has shown that by certain permissible assumptions as to the change in dissociation constants for H_2S and HS^- in extrapolating these to higher temperatures, galena may pass through a local solubility maximum on its solubility curve at about 200° C. Crystals previously deposited during falling higher temperatures are leached near this temperature, but galena is again deposited at lower temperatures. This leaching temperature is somewhat higher than the deposition temperatures surmised for the Tri-State ore deposits, but it conceivably could be modified by factors not considered in Helgeson's thermodynamic study.

Similarity to Other Hydrothermal Deposits

There is no fundamental difference between the mineralogy of the Tri-State ores and that of the simplest assemblages in many western mesothermal deposits that

are undoubtedly related genetically to igneous intrusives. As pointed out in the previous section, the resemblance includes the composition of the ore solutions, as indicated by inclusions entrapped during the crystallization of certain minerals. The resemblance also extends to the isotopic constitution of minerals deposited from the ore solutions.

A study of the variations in isotopic composition of oxygen in calcite, dolomite, and quartz in the Leadville limestone of the Gilman area, Colorado, shows that these variations have a zonal relation to the center of most intense mineralization (Engel and others, 1958). The variations are roughly zoned in the order that would be expected on theoretical grounds as due to a temperature gradient. Table 9 abstracts and generalizes the data for the Gilman area.

TABLE 9.—*Isotopic composition of oxygen in calcite, dolomite, and quartz in relation to hydrothermal zones in the Gilman area, Colorado*

[Modified from Engel, and others, (1958). δO^{18} is a measure of the abundance of O^{18} in relation to O^{16}, expressed as a per mil deviation from the $\frac{O^{18}}{O^{16}}$ ratio in a standard, as per the following definition: $\delta O^{18}=1{,}000\left(\frac{R}{Rstd}-1\right)$, where $R=\frac{O^{18}}{O^{16}}$ in mineral sample, $Rstd=\frac{O^{18}}{O^{16}}$ in mean sea water used as the oxygen isotope standard]

Zone	Rock or mineral zones	δO^{18}(°/oo) (limestone or calcite)	δO^{18}(°/oo) (dolomite)	δO^{18}(°/oo) (chert or quartz)
1	Limestone, chert, and sedimentary or diagenetic dolomite, unaltered	21.0–23.9	27.6–28.6	28.0–29.4
2	Rocks showing recrystallization, incipient hydrothermal dolomitization, or altered to "zebra rock" on periphery of the dolomite mass at Gilman, and in or bordering small low-grade sulfide bodies in outlying districts	14.6–17.8	20.1–25.8	22.2–24.2
3a	Fine-grained recrystallized dark dolomite forming the matrix of the "zebra rock," and recrystallized chert, in area of intense hydrothermal dolomitic alteration at Gilman, within 4,000 ft of ore		20.7–23.4	22.5
3b	Medium to coarsely recrystallized bands in "zebra rock" in area of intense hydrothermal dolomitic alteration at Gilman, 500–4,000 ft from ore		17.9–19.7	
4	Minerals from Gilman ore body and within 100 ft of it	13.4	16.4–17.4	16.1–17.1

At the suggestion of A. E. J. Engel, Robert N. Clayton, of the California Institute of Technology, kindly determined the isotopic composition of the oxygen and carbon in 15 dolomite and calcite samples furnished by us from the Picher field. Table 10 gives the results of these experiements. In the dolomite samples, which include both pink and gray spar, there is surprisingly small variation in the isotopic composition of both oxygen and carbon. The calcite samples reveal considerably more variation. In comparing these figures with those for the Gilman area, it must be remembered that the isotopic composition of the minerals is controlled by that of the solutions from which they were deposited, as well as by temperature. If it can be assumed that the ore solutions at Picher and Gilman were isotopically similar, then the temperatures of dolomite deposition at Picher, as judged from the oxygen isotopes, were probably comparable to those in zone 2 at Gilman, namely, the deposition temperatures of small sulfide bodies in outlying districts or of the border of the major dolomite mass during the main surge of hydrothermal dolomitization. Zone 3 at Gilman lies nearer to the hydrothermal source; but a part of its dolomite (3a) apparently records the same temperature intensity as zone 2, because this part was not recrystallized at the later higher temperature stage (3b). At the present state of knowledge the oxygen isotopic compositions can indicate only relative temperatures, but it would be reasonable to assume that the actual temperatures were somewhat above those of normal ground waters.

TABLE 10.—*Isotopic composition of oxygen and carbon in calcite and dolomite from selected samples in the Picher field*

[Determined by Robert N. Clayton. Definitions as in table 9; the carbon isotope standard is the carbon of the Cretaceous belemnite as used by Urey and others, 1951]

Specimen	Mine	Description	δO^{18} (°/oo)	δC^{13} (°/oo)
		Calcite		
1	John Beaver, south forty.	Core of clear scalenohedral crystal.	18.00	−3.68
2a	Tulsa Quapaw, north forty, K bed level.	Core of opaque, whitish calcite	21.24	−3.49
2b	do	Pale-yellow scalenohedral crystals on sample 2a.	15.31	−4.62
3	Federal Jarrett, southeast fraction.	Large scalenohedral crystal that includes marcasite needles and chalcopyrite near its surface; from barren natural cave, 400 ft long and 50 ft wide.	17.35	−1.76
4	Webber, southeast forty.	Center of prismatic crystal with ore.	15.31	−4.43
5	Robinson, east forty	Whitish calcite interstitial to porous jasperoid.	17.51	−3.64
6a	Chubb, west fraction	Clear scalenohedral core of "corn cob".	19.75	−2.86
6b	do	White porous skeletal flat rhombic calcite in outer envelope of "corn cob".	20.12	−6.08
7	Tulsa Quapaw, south fraction.	Complex crystals, dominantly scalenohedral with steep and unit positive rhombs.	21.74	−4.37
		Dolomite		
8a	Midcontinent, north fraction.	Fresh pink spar crystals	21.57	−1.43
8b	do	Gray spar with sample 8a	20.60	−1.59
9a	King Brand, northwest forty	Fresh pink spar crystals	21.32	−1.51
9b	do	Gray spar with sample 9a	20.83	−1.60
10a	Boska	Euhedral pink spar, projecting into cavity.	21.16	−1.37
10b	do	Subhedral pink spar with sample 10a in matrix of a chert breccia.	21.22	−1.37

The calcite analyses suggest some variation, with temperatures ranging between those of Gilman Zone 2 and Zone 1, the latter representing virtually normal ground-water temperatures. The "corn cob" calcite of specimen 6, which has the low-temperature isotope composition, is known from its paragenetic relations to be later than other types and it could conceivably be a cold-water phase, much younger than the ore. The crystals of specimen 7, which also are of low-temperature

isotopic composition, are of the same general type as others that have an isotopic composition indicative of higher temperature. The specimen grew on the wall of a large stope where maximum temperatures for the deposition stage would be expected to have prevailed. As calcite is the youngest mineral in the paragentic sequence, it is probable that some of it might well have been deposited at or near normal ground-water temperatures. Because of uncertainities as to the isotopic constancy of the depositing solutions, particularly in the latest stages, perhaps the only significance that can be attached to the oxygen isotopic composition of the calcite is that this composition is consistent with hydrothermal solutions and with the temperatures established by Schmidt (1962) from the study of fluid inclusions (p. 146). The isotopic compositions of calcite and dolomite cannot be paired to delineate accurate temperatures or to define the isotopic composition of the depositing solution (Clayton and Epstein, 1958, p. 365–367) because these minerals are not of simultaneous crystallization from a common parent solution.

The variations in the δC^{13} values of the Picher carbonates are consistent in that the six dolomites have a narrow range, from -1.37 to -1.60, and all the calcite of types that normally accompany ore range between -3.49 and -4.62. Of the three calcite samples outside this range, No. 3 is a coarse scalenohedral type that fills caves characteristically barren of ore, whereas the other two (6a, b) are "corn cob" calcite, possibly unrelated to ore. Until more information is available from studies in other districts in which the origin is better understood, the significance of these carbon-isotope findings in terms of physical environment during deposition will remain unknown.

In summary, the more that we learn about the Tri-State ores, the more features we find they have in common with ores of unquestioned magmatic derivation, and only the stratigraphic and structural settings are unusual. But these differences have no genetic significance. It seems preferable to assume a common origin for all rather than try to justify the unique origins ascribed to Tri-State ores in the past, when the resemblances were less known.

Conclusions

Under the preferred interpretation, the Tri-State ores are "telethermal" and may have been deposited at considerable distance from their magmatic source. Furthermore, the ore solutions may well have picked up certain constituents, such as dolomite, from the wall-rocks of the channels traversed. The later courses of the solutions probably had more of a horizontal than a vertical component of direction, as they anastomosed through the Boone Formation at the top of a carbonate rock section, under a tight shale capping.

Perhaps a structural feature, such as the Miami trough, had genetic significance only as a break that tapped an igneous source in depth. This break has been mapped in a straight course over a length of 42 miles by Weidman (1932) and Pierce and Courtier (1938) (p. 74); but only in the 7 miles or so through the Picher district is the trough axis adjacent to ore, except for a small mineralized area within a mile of the axis near Crestline, Kans., 15 miles northeast of Picher. The trough must reflect a break in the Precambrian basement. The large-scale areal pattern of ore bodies in the Picher field (pl. 1) indicates that the Miami trough is accompanied by other breaks in the basement, in a conjugate system whose directions are, respectively, parallel and nearly at right angles to the main break. The Seneca graben, a similar northeast break, has small lead-zinc deposits along it for several miles northeast from the Oklahoma-Missouri State line, but these have been of no great importance.

Ohle (1959) has pointed out the ready flexibility of a theory related to igneous activity in explaining many features of the Mississippi Valley type of ore deposits, but he calls attention to several puzzling features. These include the usual lack of igneous rock outcrops, and the broad distribution of the ores in certain districts, with the implication that either the solutions traveled a long way from the source, or the source itself was very large and gave off remarkably uniform ore emanations throughout its mass. Without directly advocating the magmatic-hydrothermal theory, he says (p. 786):

> Such igneous rocks as are known in these regions are almost all basic, without acid counterparts; most of the areas are passive tectonically and there really is no direct evidence that magmatic differentiation has occurred in them. Is it possible that these emanations stem from a different kind of igneous source, more widespread in extent, which occurs at great depth and has no especial tendency to move upward to form intrusions? * * * these deposits, if they are hydrothermal, must indeed be a product of some variant from the simple hydrothermal family tree.

Although there are no known intrusive rocks near the Picher field, evidence for an unusual type of igneous activity is widely scattered through the general area containing the Mississippi Valley type of ore deposits in western Kentucky, southern Illinois, Missouri, Arkansas, and southeastern Kansas. The known igneous features of post-Precambrian age consist of diatremes, ash beds, stocks, plugs, pipes, dikes, and sills, and these commonly contain rocks of alkalic composition (Brock and Heyl, 1961). Both felsic and mafic facies have been

described, but much of the felsic material can be interpreted as fragmental debris carried or blown out by explosive volcanic activity from the Precambrian granitic basement (see, for example, Clegg and Bradbury, 1956, p. 15). Thus, the igneous rocks of possible significance are the alkaline syenites, found only in Arkansas, and more widely distributed mafic types such as characterize the smaller igneous bodies in alkaline provinces (Kemp, J. F., in Williams, 1891, pp. 400–406).

Association of the igneous features with ore deposits is not close and is for the most part fortuitous. The two occupy similar environments of relatively mild structural deformation characteristic of the central craton. The closest of the igneous features to the Picher field is the Rose dome, in Woodson County, Kans., about 73 miles northwest of Picher, Okla. This is an unexposed intrusive laccolith(?), with an exposed granite porphyry dike, and mafic dikes or sills that were intersected in deep wells (Knight and Landes, 1932). Rose dome lies at the west end of a linear 400-mile zone along which, at irregularly spaced intervals, deep-seated igneous activity was manifested at different times between the Cambrian and Cretaceous periods (Brock and Heyl, 1961; Snyder and Gerdemann, 1965).

POSSIBLE BEARING OF ISOTOPIC CHARACTER OF LEAD ON ORIGIN

The genetic history of the lead in ore deposits is reflected in the isotopic composition of the lead, but the interpretation of this history is not easy. The initial work published by Nier (1938) started with the fact previously established that common lead is composed of four isotopes, three of which (Pb^{206}, Pb^{207}, Pb^{208}) are produced by radioactive disintegrations (of U^{238}, U^{235} (=actinium), and Th^{232}, respectively). If samples of common lead showing the lowest percentages of Pb^{206}, Pb^{207}, and Pb^{208} relative to the fourth or nonradiogenic isotope (Pb^{204}) could be considered as primeval lead that existed when the earth's crust was formed, then other common leads could be conceived as this primeval lead plus later contaminations by radiogenic lead. The contamination would, of course, have occurred from association with uranium, actinium, and thorium at some stage in the geologic history before the given sample was deposited in its present form. This contamination was indicated by Nier for samples from various localities by subtracting from the isotope abundances (relative to Pb^{204}) the three radiogenic isotopes in the proportions in which they occur in primeval lead, the excess indicating the radiogenic contamination. Later work by Patterson and others (Patterson, 1953, 1956; Patterson and others, 1953, 1955) has indicated that the small amount of lead in troilite (FeS) of iron meteorites has primitive isotopic composition consistent with the concept that it may resemble primeval earth lead.

Many of the lead deposits of the world, and particularly those of such size and grade as to be of commercial importance, have shown a parallelism of isotopic evolution that indicates that they have come from a source of worldwide extent in which the proportions of uranium, actinium, and thorium relative to lead have remained nearly constant since the outer zones of the earth were stabilized, except for changes produced by radioactive disintegration (Russell and Farquhar, 1960, p. 10–12, 66; Cannon and others, 1961). Russel and Farquhar (1960, p. 53, 62) postulate that this uniform source is in the mantle of the earth. The proportion of radiogenic increment in the lead from such an environment would then depend primarily on the time at which the lead was separated from the radioactive materials in the mantle to form the deposit. Lead with such history has isotope ratios that lie along a calculable growth curve, and has been named "ordinary" lead.

Lead from a few deposits, however, does not conform to the standard evolution. It may contain a deficiency or an excess of radiogenic lead. That with an excess has been termed "exceptional" lead by Cannon and others (1961, p. 23) and "anomalous" lead by Russell and Farquhar (1960, p. 60–62), who use this term in a restricted sense from the usage of certain other authors. Exceptional lead can be further classified into U-lead, Th-lead, and J-lead (J for Joplin), depending on whether the radiogenic excess has been derived dominantly from uranium, thorium, or both (Cannon and others, 1961). The natural boundaries between these different classes are gradational. Pb^{206} and Pb^{207} are associated in mathematically rigid proportions in any radiogenic addition, determined by (1) the ratio of U^{238} to U^{235} in primeval uranium, (2) the quite different decay rates of these two elements, and (3) the time at which the radiogenic lead was separated from the parent uranium; for there has been no natural fractionation between U^{238} and U^{235}, and at most, only slight fractionation between Pb^{206} and Pb^{207}.

Of about 20 localities throughout the world that were investigated by Nier and his coworkers (Nier, 1938; Nier and others, 1941), the samples of galena from Joplin had the highest radiogenic contamination, well in excess of the proportions expected from ordinary evolution. This could mean that (1) the original (magmatic?) source from which the lead was derived was exceptionally rich in uranium and thorium so that accretions of radiogenic lead had been relatively large; (2) the source may not have been so rich, but the

accretions occurred over a longer period of time before the lead was separated from its radioactive parent and deposited; or (3) the lead in the mineralizing solutions may have originally been ordinary, but it picked up additional radiogenic lead from the rocks traversed after leaving the source.

Later isotopic studies by several workers have amply corroborated the large radiogenic contamination for Tri-State lead and for Mississippi Valley type lead in general (see Russell and Farquhar, 1960, p. 117–223; Cannon and others, 1961, p. 24–29). However, lead from other areas also shows excess contamination to varying degrees. Thus, certain of the veins at Sudbury and in the Thunder Bay region, Ontario, contain extreme examples of anomalous lead (Russell and Farquhar, 1960, p. 71–74); and in west-central New Mexico deposits of several types but predominantly veins and breccia zones contain anomalous lead which occurs along both fringes of a tectonic belt, the center line of which is characterized by the presence of ordinary lead (Slawson and Austin, 1962). The geologic environment of these several districts shows little resemblance to that of Mississippi Valley deposits. The country rocks are varied, including limestone, volcanic rocks, and Precambrian granites and gneisses; and whereas some districts contain intrusive igneous rocks, others do not.

The radiogenic component of exceptional (J) lead in the Tri-State region is an addition to a predominating content of ordinary lead. Hence, on a triangular composition diagram having Pb^{206}, Pb^{207}, and Pb^{208} at the corners, the curve for Tri-State lead compositions is displaced from the ordinary lead curve in the direction of more radiogenic composition, but the displacement is slight in comparison with the total possible range of isotopic composition (Cannon and others, 1961, figs. 3, 18).

Anomalous—or exceptional—leads, whether from Mississippi Valley or other types of ore deposits, are characterized by a relatively high degree of isotopic variation between different samples within a given district (Russell and Farquhar, 1960, p. 60–66; Kulp and others, 1956; Cannon and others, 1961, p. 24). Commonly, the variations can be shown to be related along a mixing straight-line graph between ordinary lead and radiogenic lead having a relatively fixed ratio of isotopes (Russell and Farquhar, 1960, p. 63–65, 71–85; Cannon and others, 1961). The variation may even extend to different parts of a single galena crystal. Cannon and others (1963a, b) have described such a crystal from the [West] Netta mine at Picher, in which successive zones become progressively more radiogenic from the core outward, except for a surface film whose radiogenic content recedes slightly in comparison to the next underlying zone.

Russell and Farquhar (1960, p. 61–66) conceive that lead which now contains an excess of radiogenic lead left the original source in the earth's mantle at a time of orogeny, but instead of being deposited promptly in permanent form, it tarried for a short or long period in crustal rocks having relatively much higher contents of uranium and thorium than the mantle, and was finally deposited, or redeposited, in the present ore deposits. Remobilization of the lead from earlier deposits is particularly stressed. The excess radiogenic lead was dissolved by the lead-bearing solutions during their migration through the surface rocks, "during the processes by which the lead minerals were concentrated and deposited" (Russell and Farquhar, 1960, p. 78). These surface rocks are of varied types. In at least one example derivation of the anomalous lead deposits by secondary concentration from dolomitic sediments is apparently accepted (Russell and Farquhar, 1960, p. 82). Cannon and others (1963b) have presumed that the lead forming the isotopically zoned crystal from the Netta mine at Picher was leached from the Paleozoic sedimentary rocks of the general area and redeposited in the crystal during a long period, equivalent to late Paleozoic plus Mesozoic time.

It seems to us, however, that the radiogenic contamination of Tri-State lead ores more probably was derived from granitic and related felsic rocks of the Precambrian basement that were locally melted up by basaltic magma from deep-seated sources and incorporated into a hybrid magma. Precambrian rocks in the Picher area, to judge from intercepts in deep wells and from the nearest outcrops at Spavinaw, Okla., are largely granite. On the crest of the Ozark uplift where Precambrian rocks crop out, these consist mainly of granites, felsites, porphyries, and porphyritic breccias, but locally include iron ores and ferruginous slates (Van Hise and Leith, 1909, p. 740–741). Granitic rocks have a higher content of uranium and thorium and a higher ratio of these elements to lead than basaltic and related rocks (Patterson and others, 1955; Larsen and Phair, 1954; Adams, 1954). The uranium and thorium and their decay products, including radiogenic lead, are particularly concentrated in such accessory minerals as zircon, sphene, apatite, allanite, monazite, and xenotime, but they also occur to a notable degree in the mineral interstices and intracrystalline fractures where they are most readily soluble (Tilton and others, 1955). Incorporation of such rocks into primary basaltic magma, which most nearly fulfills the requirement of a stable uniform environment for the evolution of ordinary lead, would

be equivalent to mixing of ordinary lead with a radiogenic lead of fairly uniform composition. This composition would be determined by fairly uniform average ratios of uranium and thorium to each other and to lead in the granite and by the length of time that the granite has existed as a differentiate of the earth's crustal layers. A hybrid magma would contain lead of varying isotopic composition, depending on the amount of granite absorbed; and the radiogenic contamination could well increase rapidly with time as additional granite was absorbed. At the same time, the leavening effect of the ordinary lead in a magma evolving by assimilation of invaded rocks in a nearly closed system would insure that the radiogenic additions would be blended in the total lead system to produce an isotopic composition not greatly different from that of ordinary lead. Such an igneous evolution would preclude abrupt and large variations toward highly radiogenic composition that would be expected of evolution in any aqueous system within the ground, where there would be little chance for blending the radiogenic additions with the total preexisting lead. If lead were given off in hydrothermal solutions from such a magma, this lead might well show the evolution to more radiogenic composition recorded in the successive zones of the crystal described by Cannon and others (1963a, b).

Hybrid magmas of the origin here suggested are considered by Turner and Verhoogen (1960, p. 382–388) to be a permissible source of granitic rock bodies, though for the bulk of large granitic bodies they favor an origin by partial or nearly complete fusion at depth of preexisting granite or varied rocks of similar average composition. The isotopic character of the lead is probably also consistent with an igneous origin from granitic magmas derived in depth by such differential fusion, without involvement of a basaltic component. Part of the radiogenic lead in the deep-seated source rock may be locked in accessory minerals that are refractory under the conditions of partial fusion. Variability in the amount of radiogenic lead in the fluid magma could thus depend on the degree to which, or the speed with which, such material was resorbed into the fluid magma from whence it could be expelled in ore solutions. As the heat for igneous fusion is probably in large part from radioactive disintegrations (Turner and Verhoogen, 1960, p. 436–447), probably that part of the Precambriam crust richest in uranium, thorium, and potassium would be the most likely locus for a molten magma.

That at least parts of the central craton are underlain by rocks containing an excess of radioactive materials is attested by the helium fields in parts of the midcontinent oil and gas field, by the uranium content of some oils, and by the presence of small radium-bearing deposits in the vicinity of certain domes ascribed to igneous intrusion.

Helium is a recurring product of the disintegration of uranium and thorium and their daughter radioactive elements, ending with the isotopes of lead which are left as stable residua. Although several theories have been advanced to explain the presence of helium in natural gas, the most plausible is that the helium was derived from the disintegration in depth of radioactive materials, ultimately derived from uranium and thorium (Rogers, 1921, p. 60–68; Gott and Hill, 1953, p. 118–119). A part of the helium is retained in the host rock, but some of it escapes and is then subject to the same laws governing accumulation as other natural gases. Helium-rich gas fields in the United States are chiefly in the midcontinent region, west and southwest of the Ozark dome (Rogers, 1921), where the stratigraphic and structural conditions were particularly favorable for entrapment. Some of the crude oils in parts of this general area are relatively rich in uranium, which in some of its compounds is soluble in petroleum (Hyden, 1961). That the lead-zinc fields in and bordering the Ozark dome lack any traces of helium or uranium has no significance, for the special stratigraphic and structural conditions essential for entrapment of gas and oil do not exist there. However, it is quite probable that as much helium may have been evolved there as in the known entrapment areas, but it has escaped into the atmosphere.

The occurrence of radium in certain precipitates from oil-well fluids in southeastern Kansas, centering around Butler County, points to the former presence in the oil-bearing strata of appreciable uranium which was largely leached out during the extraction of the oil (Gott and Hill, 1953). Gott and Hill believe that the uranium is of hydrothermal origin, introduced into Pennsylvanian and older strata from a Pennsylvanian or later intrusive. The localization of uranium again suggests an excess of this element in the subsurface rocks of the area.

The most complete sequence of post-Precambrian igneous rocks in the Mississippi Valley region is exposed in Arkansas, where early study revealed the alkaline nature of both the larger stocks and the associated dike rocks (Williams, 1891). Many of the rock analyses reported (Williams, 1891, p. 70, 81, 88, 135, 238, 263, 266, 276, 287, 399; Gordon and others, 1958, p. 61; Erickson and Blade, 1963, p. 9, 12, 13, 16, 27, 33, 41, 42, 45, 46, 48; Washington, 1901) are as rich in potassium as the "potash-rich basic volcanic rocks"

Ouachita orogeny at the end of the Pennsylvanian is unlikely. Although much of the structural setting may have been completed by the end of this orogeny, slight additional adjustments prior to mineralization could have opened the ground to the ore solutions.

EXPLORATION AND OUTLOOK

The known ore deposits of the Picher field, and the Tri-State region as a whole, are relatively simple in gross features and geologic environment. They are chiefly tabular bodies of easily recognized ore minerals (sphalerite and galena) and gangue minerals (dolomite and jasperoid) that replace favorable limestone beds in the Boone Formation. This formation is nearly flat lying throughout the region, and it either crops out or is buried only a few hundred feet beneath a gently rolling land surface of low relief. These relations made the deposits amenable to low-cost exploration by churn drilling, and for many decades the mining industry of the region depended upon drilling in geometric or haphazard patterns, with little or no geologic guidance, to find new deposits.

In the 1920's, after the broad outline of the Picher field had been established and some mines were approaching exhaustion, the usefulness of detailed stratigraphic and structural studies in guiding exploration was demonstrated by Fowler and Lyden (1932). They divided the Boone Formation into a dozen units, which permitted detailed correlation of beds, the recognition of original variations in thickness and lithologies plus those resulting from ore mineralization, and close control on the vertical position of the beds. The vertical or structural information obtained from churn drilling proved useful in planning mine development as well as recognizing the minor flexing of beds that seems to have influenced ore localization. Identification from drill cuttings of partial leaching of the limestone and associated thinning of the beds with increased percentage of chert, and the presence of clay, shale, cavities, and traces of gangue minerals, permitted the recognition of altered halos around ore bodies, thereby justifying the drilling of offset holes nearby to find ore. Geologic mapping in the mines also showed the general parallelism between fracture zones and the trends of elongate ore bodies, which proved useful in planning exploration and mine operations. Geologic guidance of exploratory drilling and mining was soon adopted by many companies working in this field.

Preliminary maps showing dolomitized areas and zonal relationships in parts of the Picher field were published by the Geological Survey during World War II (McKnight and others, 1944). These maps and the zoning concept were useful in explaining the geologic habits of many narrow ore runs, their tendency to form complete circles, and the fairly common vertical juxtaposition of some ore bodies in M and K beds. (See maps of the Tulsa Quapaw mine shown on plate 11 and numerous other mines shown on plates 5–10.) This information guided exploratory drilling where zonal relations were established or suspected, and it encouraged continued mining of narrow and commonly rich ore runs beyond working faces where the grade of the ore was locally too low for profitable mining. It also encouraged exploring for ore runs in M bed by drilling downward from the edges of K bed stopes.

Although the peak production from the Picher field was reached in 1925, and some ore bodies were virtually mined out by that time, application of these several geologic guides in subsequent years yielded enough ore discoveries to sustain production at a moderately high level in most years until the late 1950's, when production declined drastically, owing mainly to the depletion of reserves. Some production will continue from known ore bodies, but the known reserves will not support continued production at a high level. Continued exploration by drilling from mined-out stopes for new ore bodies or extensions of known bodies in lower or higher stratigraphic levels, especially drilling for M bed ore runs from K bed stopes, will probably yield new reserves, but not in large amounts, for many of the seemingly favorable targets suggested by study of plates 5–10 have been tested already. Continued exploration by drilling from the surface, especially on the fringes of the field, might even find some new deposits in the Boone Formation, but as the entire area of the Picher field and bordering country has been rather intensively drilled, the chances of finding major ore bodies are probably slim.

Although ore production from the Boone Formation in the Picher field probably will never again be very large, this field and the entire Tri-State region will continue to receive some exploration attention by the mining industry because of the philosophy of "hunting for elephants in elephant country." A sound appraisal of the chances of finding major deposits at a lower stratigraphic horizon in the field, or a new mining district in a nearby area, depends upon interpretive and permissive geology—does the interpretation of the origin of the known deposits and the geologic features that localize them permit the concept that ore-bearing solutions passed through a given target area and found favorable host rock?

The known deposits are thought to have formed from ascending hydrothermal solutions, probably from

an igneous source; the gangue and ore minerals replaced beds of favorable limestone at places where the rock was prepared by fracturing due to minor flexing, partial solution, and associated collapse. A few general considerations of possible target areas are discussed below.

In the Picher field, Paleozoic beds below the Boone Formation consist of several formations of Cambrian and Ordovician age (table 1). The Chattanooga Shale of Devonian and Mississippian age is present just south of the field. These units are known only by intersection in a few deep drill holes, and not all of them are present in each hole. This drilling has shown that the topographic relief on the surface of Precambrian rocks is as much as several hundred feet in the Picher field, and on the topographic highs of this surface some of the older formations are missing, probably due to nondeposition. Thus, the total thickness of the lower Paleozoic beds probably ranges from about 700 or 800 feet on the highs to 1,200 or 1,300 feet in the low places. The Chattanooga Shale is only a few feet thick in holes along the south edge of the Picher field, it is generally absent or thin and discontinuous within the field, and it is not reported in wells to the north.

The Cambrian and Ordovician beds under the Picher field are composed dominantly of dolomite with some chert; a little sandstone and shale are present in some units. Dolomite of this general type is host to lead and zinc deposits in many parts of the world, and in southeastern Missouri, beds of dolomite of the same general ages and lithologies are host to the highly productive lead deposits. Beneath the Picher field the Cambrian and Ordovician beds have undoubtedly been flexed in places, and probably they have been somewhat broken by fracturing. In addition, almost certainly they have been subjected to some solution and collapse, for the conspicuous slump pipes shown by contours on the top of Grand Falls Chert Member (N bed) of the Boone Formation (pls. 5–10) are thought to be in part rooted in these beds. And certainly, if the solutions that formed the ores in the Boone ascended from depth, as is supposed, they would have had to pass through these beds. Thus, these Cambrian and Ordovician beds must be considered potentially favorable or at least permissible host rocks from the standpoint of lithology, structure, and the passage of ore solutions.

On the other hand, the absence of Chattanooga Shale over most of the field might possibly be a unfavorable factor. Almost everywhere in the Picher field, beds of the Boone Formation and Chester Series are overlain by shale of the Krebs, and this cover was probably even more extensive at the time of ore mineralization. Thus, this shale would have formed a nearly impervious blanket. The resulting retardation of the ascending solutions might have been a factor in the localization of ore in the Boone Formation, and if this is true, perhaps the solutions passed through the Cambrian and Ordovician beds too freely to form deposits. South of the Picher field, however, the Chattanooga Shale is present, and it is possibly thick enough (several tens of feet) to have retarded ascending solutions; so perhaps these Cambrian and Ordovician beds are more favorable at places south of the Picher field.

In considering exploration in areas nearby the Picher field, additional factors merit attention. A conspicuous northeast-trending fracture, the Miami trough, crosses the west side of the Picher field and extends miles beyond it to the northeast and southwest. Within the field this trough seems to have influenced the localization and shape of ore bodies in places but not everywhere along it. This fracture is assumed to extend deeply into the underlying basement rocks, and at depth it could have been a major channelway for ascending solutions. A small mineralized area close to the trough near Crestline, Kans., and small deposits along a similar structure—the Seneca graben—in southwestern Missouri, perhaps can be viewed as evidence to support this concept. Thus, places along conspicuous structures that might extend to depth and be solution channelways deserve consideration as sites for exploration.

Even so, in the Picher field the known deposits extend several miles southeast and northwest of the Miami trough, so far that it hardly seems logical that this trough was the only feeding channelway at depth; rather, it seems more reasonable to assume that other deep-seated channelways also fed solutions upward even though they have no surface structural expression like that of the Miami trough. Thus, ground that lacks an exposed conspicuous structural feature probably should not be eliminated from exploration on this basis alone, but perhaps should be given lower priority.

Any extensive exploration program should test Cambrian and Ordovician formations as well as Mississippian beds. Such a program will be costly, and it might be slow in yielding satisfactory results. Perhaps it will not be undertaken until the world reserves of zinc ore are less adequate to supply requirements than in the mid–1960's. But sooner or later, conditions will probably justify extensive exploration; when this is done, all applicable prospecting techniques should be used.

In recent years geophysical prospecting techniques have advanced rapidly, and increasingly useful results should be expected in future years. Helpful information that might be obtainable from geophysics include

some of the following: The presence of deep-seated igneous bodies that might have supplied the ore-bearing solutions; configuration on the basement topography and the recognition of major fracture systems in the Precambrian rocks; areas of deformation, solution, and compaction in Paleozoic rocks; and, hopefully, recognition of masses of gangue and ore minerals.

Because of the near-surface deposits in the field, and the contamination due to mining and milling activities, surface geochemical prospecting techniques cannot be used here to appraise the possibility of more deeply buried deposits. However, they might be useful elsewhere in deep as well as shallow prospecting. And perhaps those techniques can be applied in the study of drill cuttings or cores to yield more information than would come from visual examination of recovered rock samples. In addition to the zinc, lead, and copper, small amounts of arsenic are present in the ores (p. 147), and it might form detectable halos. Mercury concentrations in 10 samples of sphalerite tested range from 120 to 1,000 parts per billion. Although these amounts are not particularly high, at the maximum they do represent perhaps a 20-fold increase over crustal abundance; and if anomalous mercury is present in the country rock near deposits, halos might be detectable.

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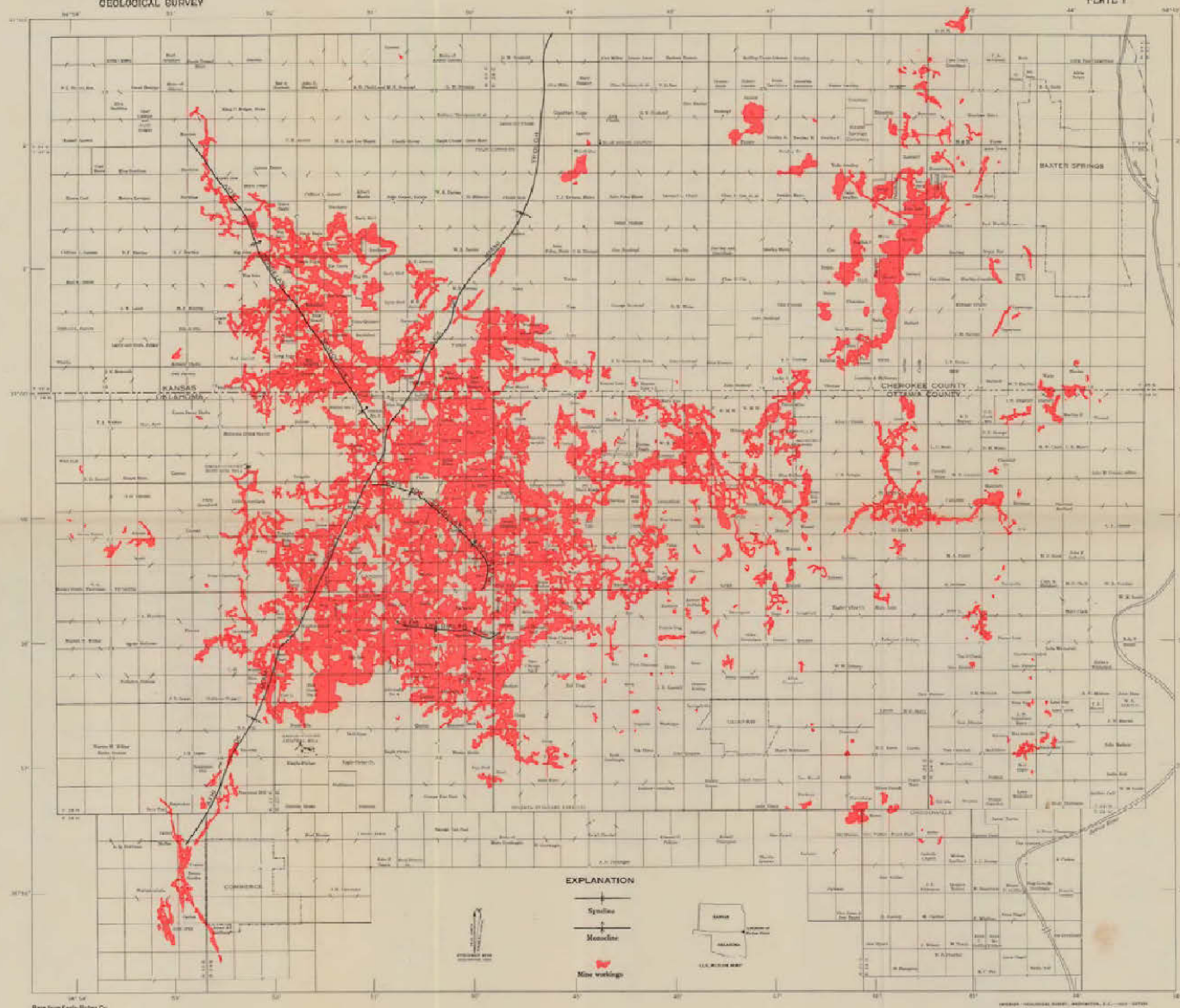
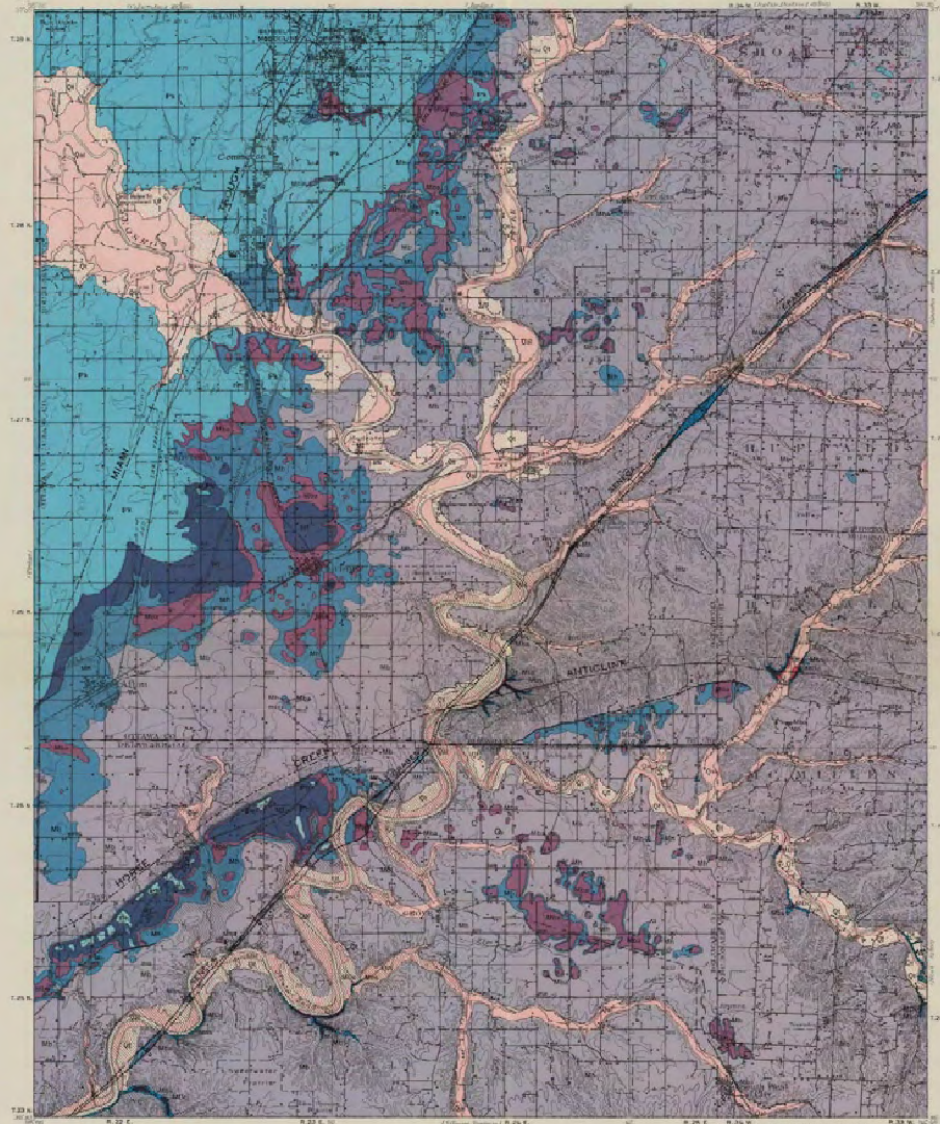
PLATE I



Base from Eagle-Picher Co.

MAP SHOWING TOWNS AND MINE WORKINGS IN THE MAIN PART OF THE PICHER FIELD OKLAHOMA AND KANSAS



EXPLANATION

SEDIMENTARY ROCKS

		Alluvium	Brown gravels, sand, and clay
	Qt	Alluvial-terrace deposits	Brown gravels, sand and clay on higher levels along valley sides
		UNCONFORMITY	
	Pk	Krebs Group	Black to gray fissile shale with a few siltstone and brown sandstone beds; sparse clay concretions. Black fossiliferous limestone, and thin coal partings
		UNCONFORMITY	
	Ph	Hale Formation	Massive to thick-carbonaceous sandstone, dark shale, and fine argillaceous fossiliferous limestone
		UNCONFORMITY	
	Mf	Fayetteville Shale	Black, thinly-bedded and laminated shale or fissile shale; minor, dense, argillaceous limestone
	Mbs	Batesville Sandstone	Gray crinoidal limestone, commonly oolitic; buff sandstone and green shale
	Mh	Hindsville Limestone	Gray crinoidal limestone, locally oolitic; sparse sandstone and green shale
		UNCONFORMITY	
	Mq	Quapaw Limestone	Gray crinoidal limestone, exposed at one locality along Spring River southeast of Quapaw
	Mb, Mm	Boone Formation	Mb, crinoidal and fine-grained limestone with oolitic and glauconitic beds; much bedded and nodular chert and sandstone Mm, St. Joe Limestone Member, gray to pink crinoidal limestone in thinly bedded
		UNCONFORMITY	
	Mc	Chattanooga Shale	Black fissile shale; thin coarse-grained white sandstone at base, locally
		UNCONFORMITY	
	Ce	Cotter Dolomite	Gray to brown dolomite, sparingly sandy, minor chert

Contact
Dashed where approximately located

Fault
Dashed where inferred; dotted where concealed. U, upthrown side; D, downthrown side

Anticline
Showing crestline; dashed where inferred

Syncline
Showing troughline; dashed where inferred

Monocline
Showing bend of strata

Area submerged by Lake O' the Cherokees

Pennsylvanian and Mississippian, undivided	Krebs Group, Fayetteville (?) Shale, Batesville Sandstone, Hindsville Limestone, and Boone Formation, undivided, in Seneca graben near Seneca and Racine	MISSISSIPPIAN, AND PENNSYLVANIAN

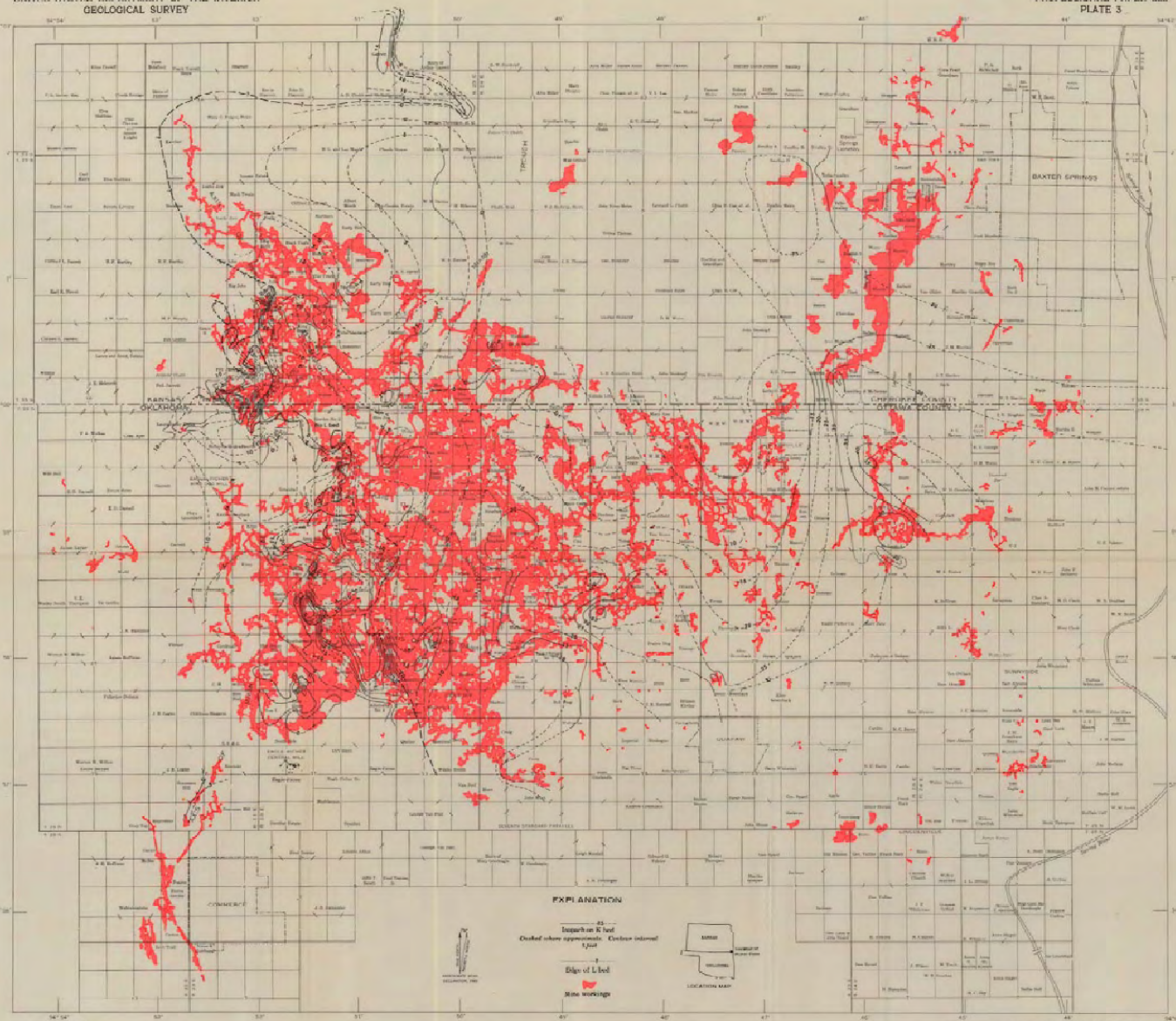
QUATERNARY
PENNSYLVANIAN
MISSISSIPPIAN
CARBONIFEROUS
DEVONIAN AND MISSISSIPPIAN
ORDOVICIAN

GEOLOGIC MAP OF THE WYANDOTTE QUADRANGLE, OKLAHOMA, MISSOURI, AND KANSAS

SCALE 1:125,000

1 0 1 2 3 4 5 6 7 8 10 MILES
1 0 1 2 3 4 5 6 8 10 KILOMETERS

CONTOUR INTERVAL 100 FEET
DATUM IS MEAN SEA LEVEL



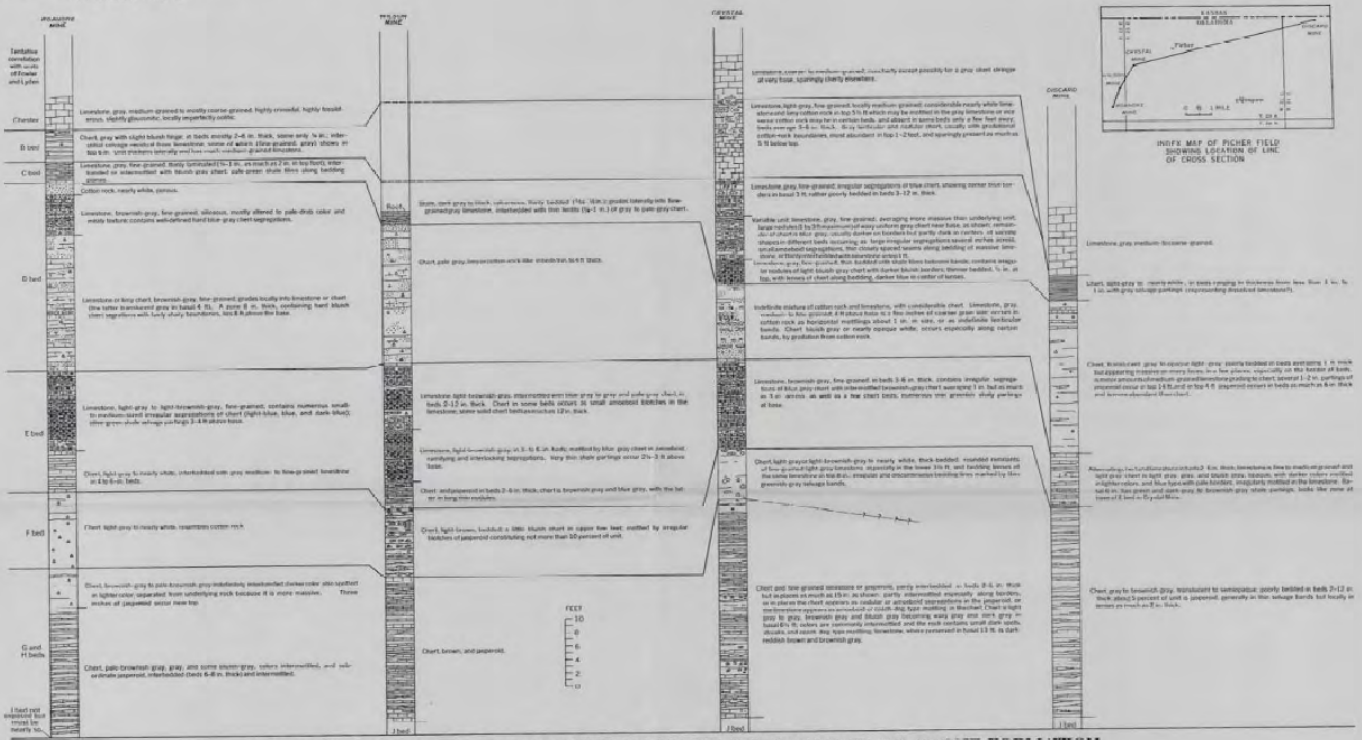
Base from Kaiser-Pitcher Co.
1:14,400, 1958

ISOPACH MAP OF K BED AND APPROXIMATE WEST LIMIT OF L BED IN THE PICHER FIELD OKLAHOMA AND KANSAS

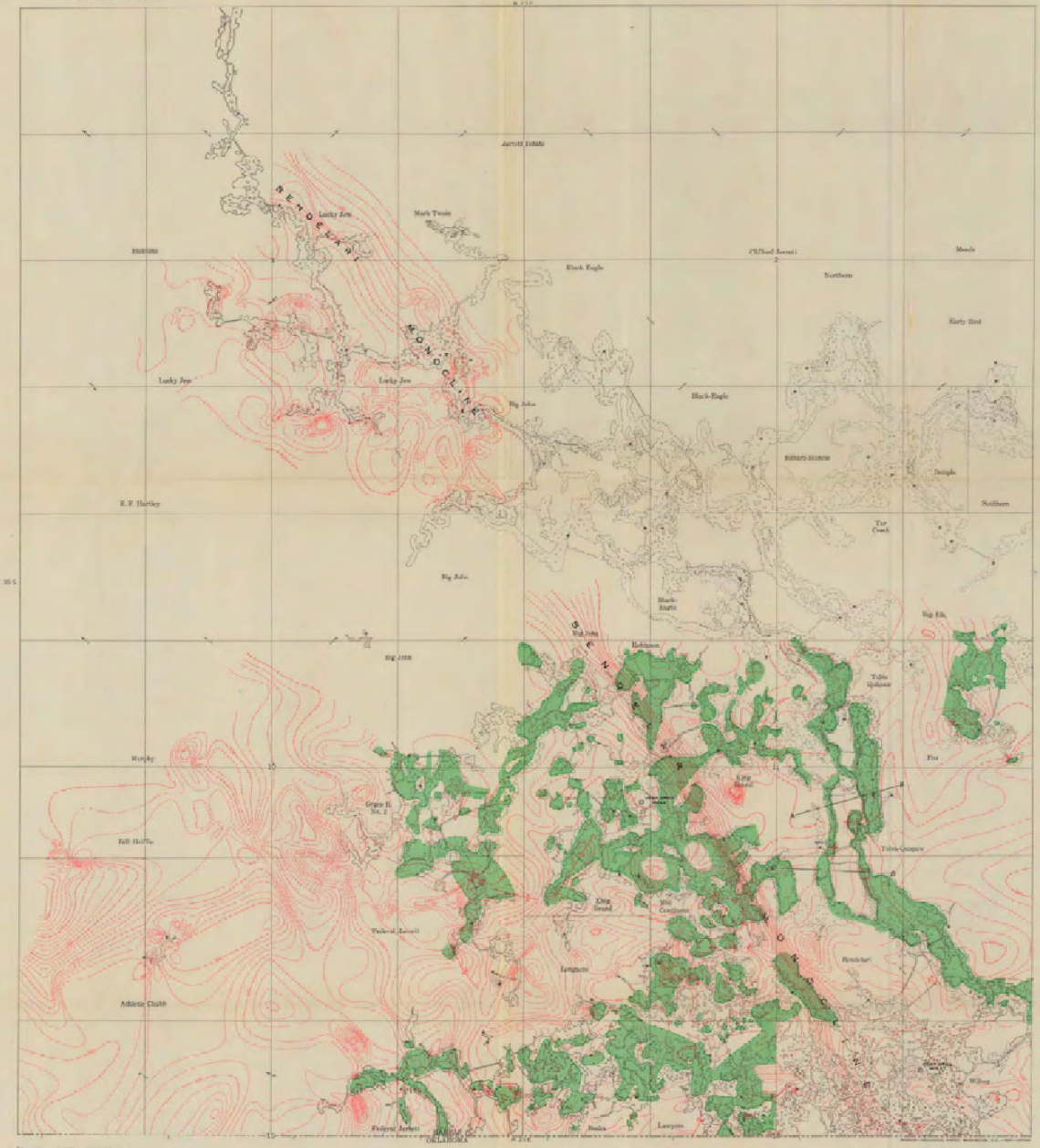
SCALE 1:31,680

1 0 1 MILE

1 0 1 KILOMETER



SECTIONS AND CORRELATION OF THE MOCCASIN BEND MEMBER OF BOONE FORMATION IN THE ROANOKE, WILSON, CRYSTAL, AND DISCARD MINES, OKLAHOMA



MAP SHOWING STRUCTURAL GEOLOGY AND DOLOMITIZED AREAS IN PART OF THE PICHER ZINC-LEAD FIELD, OKLAHOMA AND KANSAS; NORTHWEST SHEET

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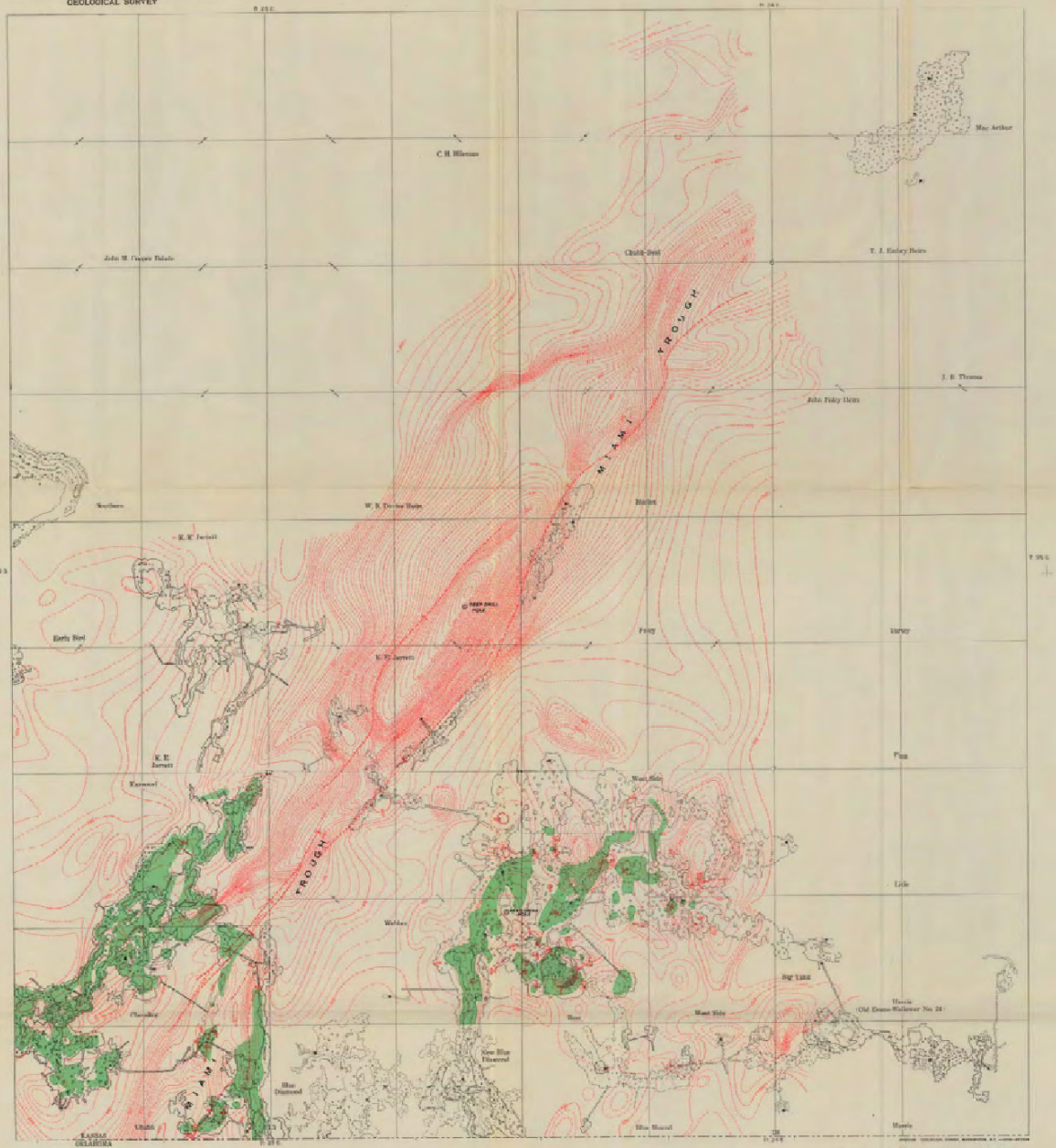
GEOLOGICAL SURVEY

PROFESSIONAL PAPER 588
PLATE 6



EXPLANATION

Unmineralized areas in M and higher beds

Fault

Sharp pipe

Elevation contour

Workings in Chester strata and E bed of Fowler and Lyden (1933)

Workings in G and H beds (Moorsale Reed Member of Boone Formation)

Workings in K bed (Baxter Springs Member of Boone Formation)

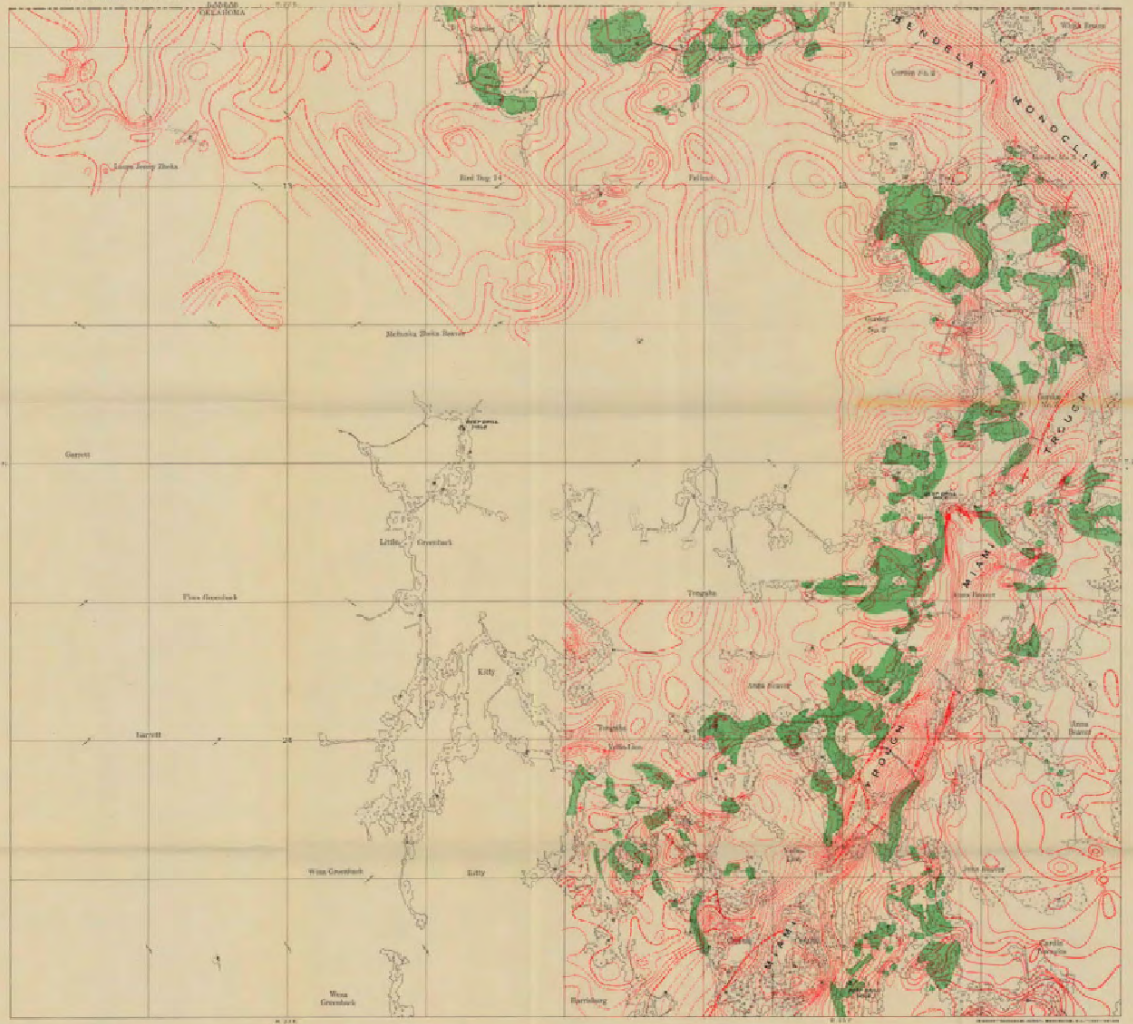
Workings in M bed (Joplin Member of Boone Formation)

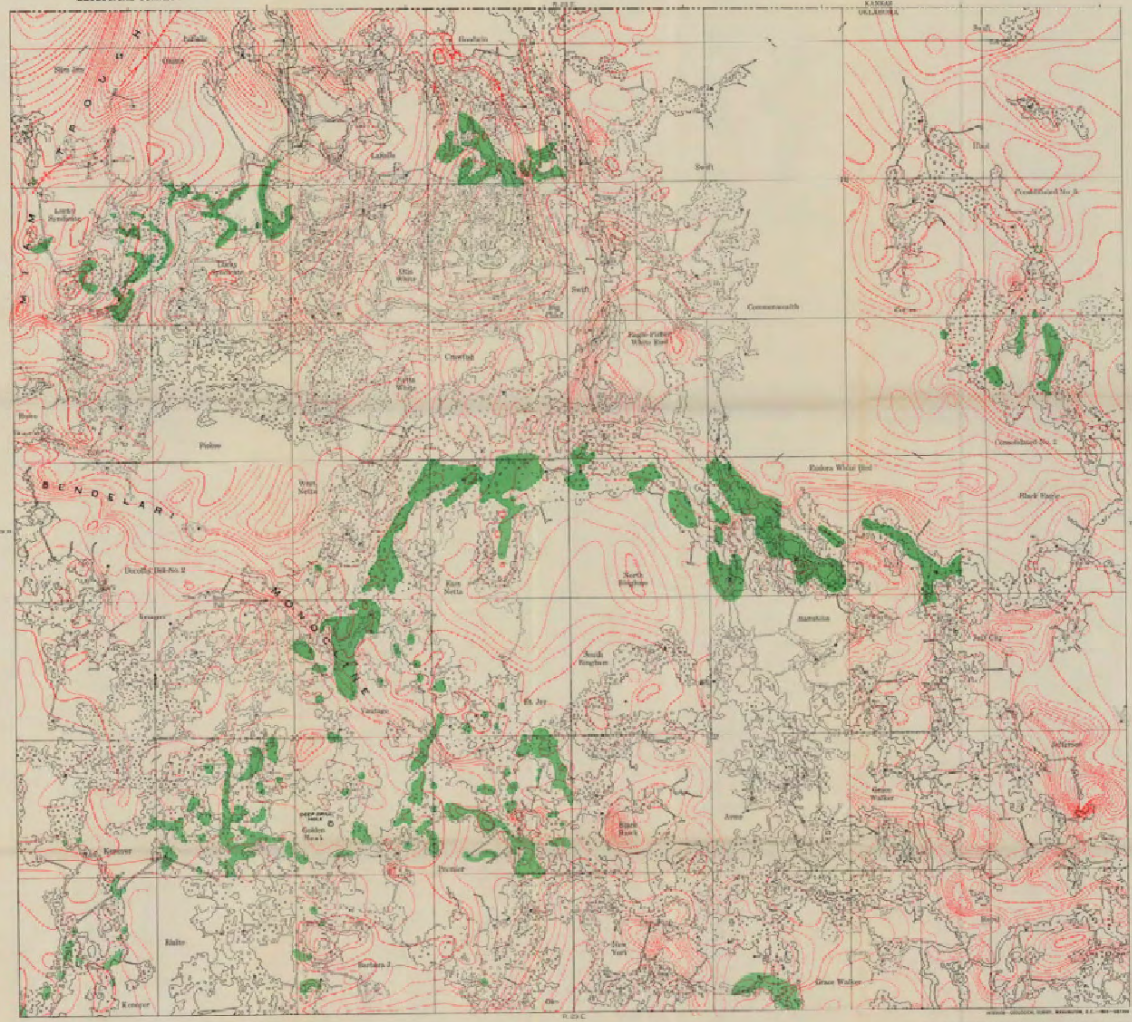
Workings in N bed "chert group" (Grand Falls Chert Member of Boone Formation)

Property line

INDEX MAP

MAP SHOWING STRUCTURAL GEOLOGY AND DOLOMITIZED AREAS IN PART OF THE PICHER ZINC-LEAD FIELD, OKLAHOMA AND KANSAS; NORTHEAST SHEET


KANSAS
OKLAHOMA
BENDELARI MONOCLINE
MIAMI TROUGH
Picher
Cardin
Quapaw
Commerce
Miami
Peoria
Wyandotte
Ottawa
Delaware
Nowata
Craig
Rogers
Mayes
Tulsa
Creek
Pawnee
Osage
Washington
Nowata
Craig
Ottawa
Delaware
Mayes
Rogers
Tulsa
Wagoner
Creek
Pawnee
Osage
Washington
Garnett
Little Greenback
Flora Greenback
Kitty
Barrett
Wren Greenback
Tergale
Lone Elm
Lone Island
Lone Beaver
Lone Island
Lone Beaver
Lone Elm
Lone Island
Lone Beaver
Spergenberg



EXPLANATION

Uneconomical areas in M and higher beds

Boundary dashed where inferred or concealed; queried, mainly to correspond where part more economical

Fault

Dashed where inferred. Ball and bar on downthrown side

Sheep pipe

Dashed where inferred. Ball and bar on downthrown side

Structure contours

Drawn on top of Grand Falls Chert Member of Boone Formation; datum is sea level. Contour interval 5 feet. Dashed where inferred

Shaft

Workings in N bed (Short Springs Member of Boone Formation)

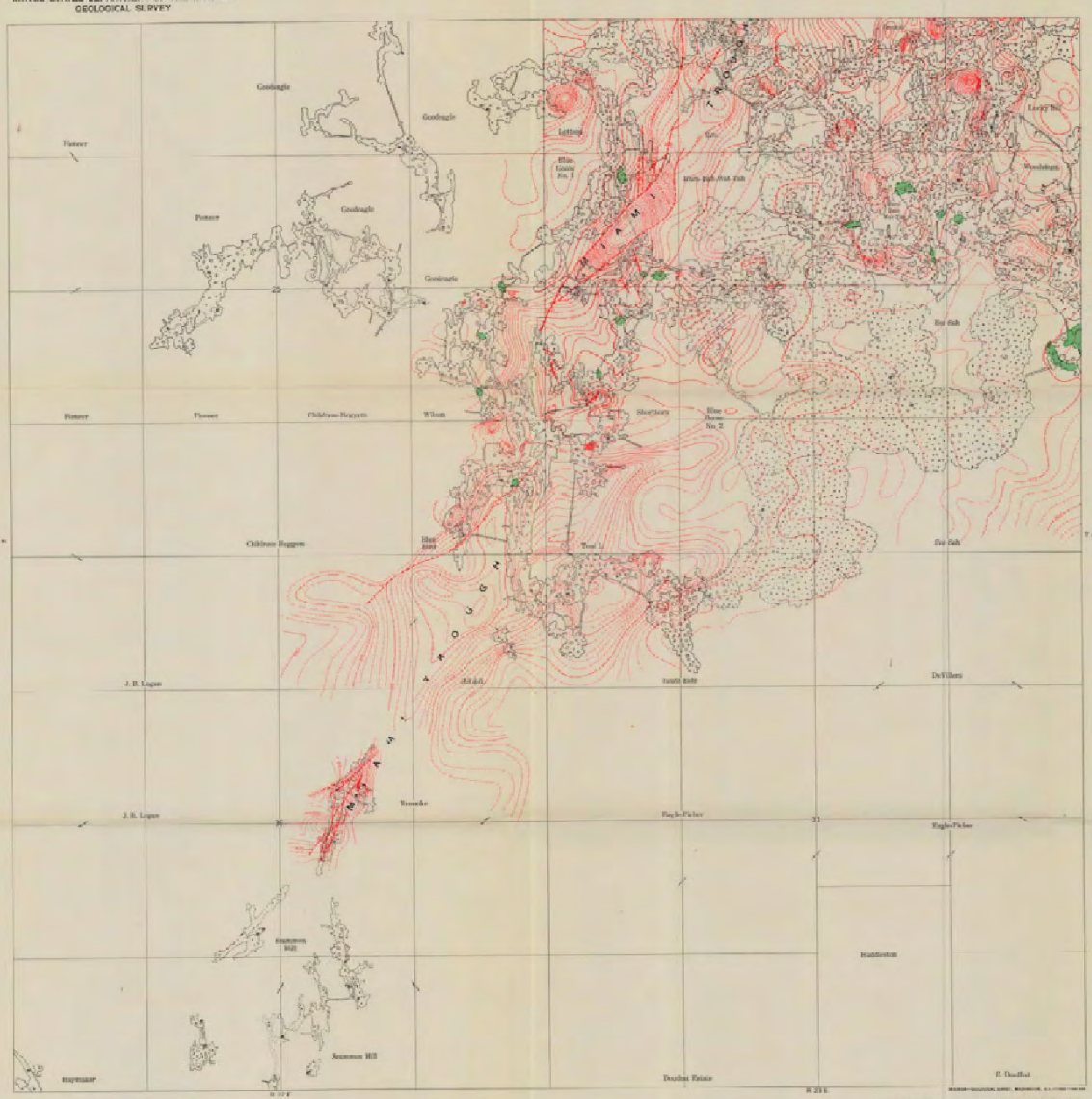
Workings in M bed (Joplin Member of Boone Formation)

Workings in N bed "chert ground" (Grand Falls Chert Member of Boone Formation)

Property tie

INDEX MAP

MAP SHOWING STRUCTURAL GEOLOGY AND DOLOMITIZED AREAS IN PART OF THE PICHER ZINC-LEAD FIELD, OKLAHOMA AND KANSAS, EAST-CENTRAL SHEET



EXPLANATION

Dolomitized areas in M and higher beds

Boundary shaded where inferred or generalized; queried, where its continuation appears rather indicated and unmarked

Fault

Dashed where inferred; D, upthrown side; U, downthrown side

Drainage gaps

Dashed where inferred; D, upthrown side; U, downthrown side

Structure contours

Drawn on top of Grand Falls Chert Member of Boone Formation, equals top of M bed of Reeds and Lyden (1933). Dashed where inferred. Hachures indicate closed basin; only structural contour in a succession downward opposite to landward. Contour interval 5 feet. Datum is mean sea level

Workings in Chester strata and in beds as low as Fowler and Lyden (1933) M bed to be Warsaw Reed Member of Boone Formation

Workings in G and H beds (Saccharoidal Reed Member of Boone Formation)

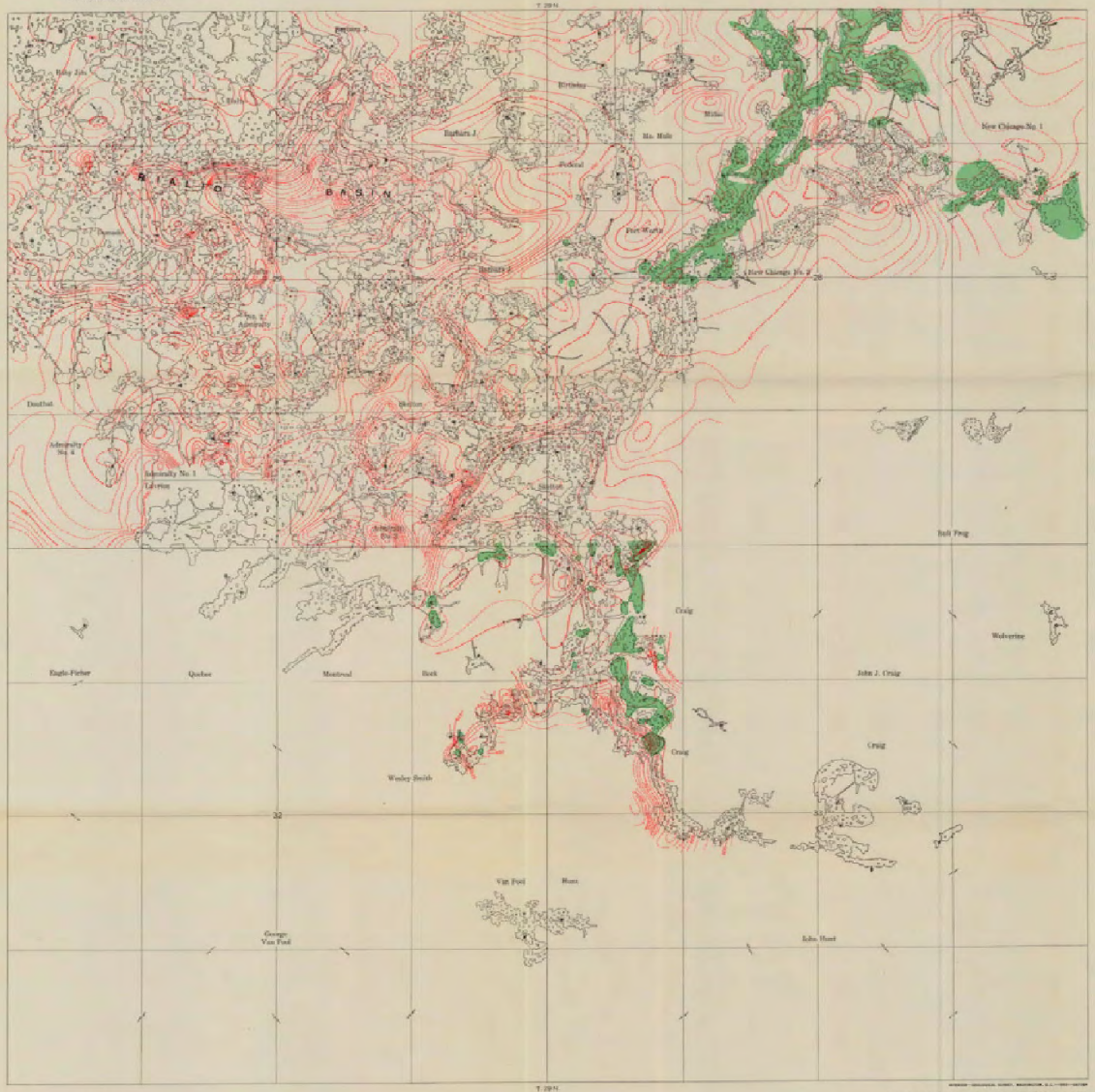
Workings in K bed (Baxter Springs Member of Boone Formation)

Workings in M bed (Joplin Member of Boone Formation)

Workings in N bed "short ground" (Grand Falls Chert Member of Boone Formation)

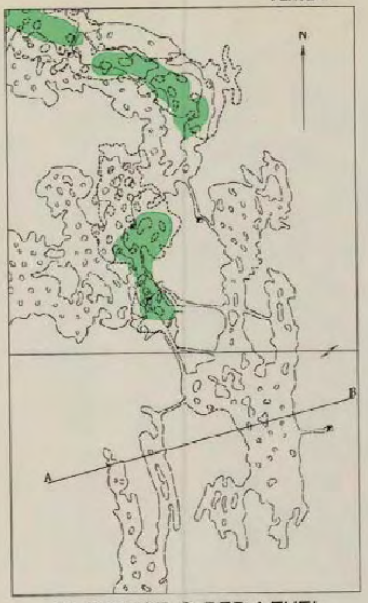
Property line

MAP SHOWING STRUCTURAL GEOLOGY AND DOLOMITIZED AREAS IN PART OF THE PICHER ZINC-LEAD FIELD, OKLAHOMA AND KANSAS; SOUTHWEST SHEET

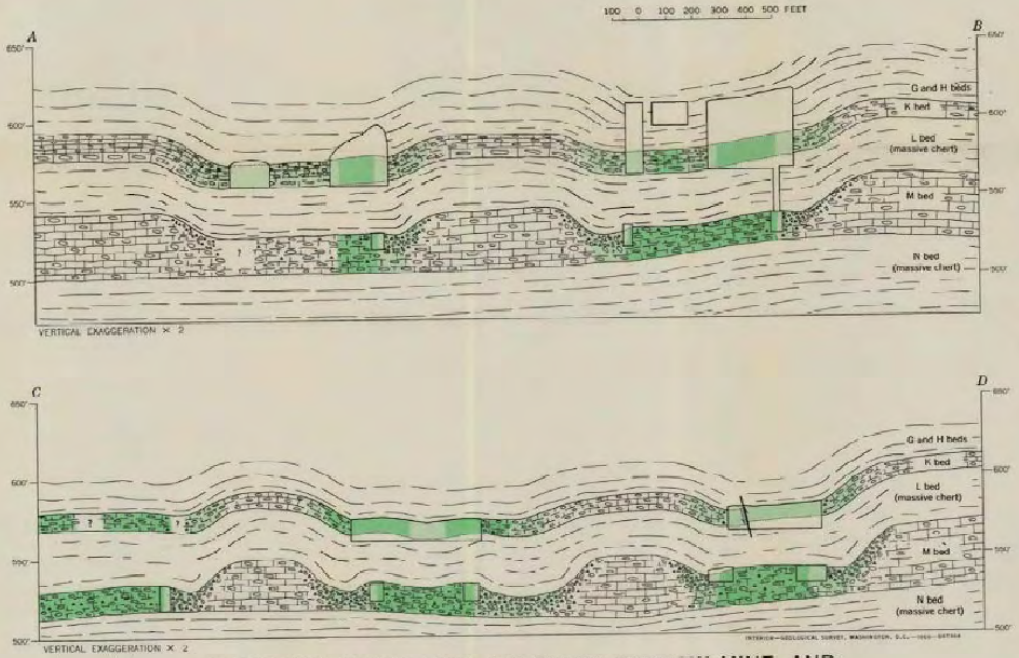

T. 29 N.
Baby Ada
Bartlett 3
Bartlett 1
Bartlett 2
Bartlett 4
Bartlett 5
Bartlett 6
Bartlett 7
Bartlett 8
Bartlett 9
Bartlett 10
Bartlett 11
Bartlett 12
Bartlett 13
Bartlett 14
Bartlett 15
Bartlett 16
Bartlett 17
Bartlett 18
Bartlett 19
Bartlett 20
Bartlett 21
Bartlett 22
Bartlett 23
Bartlett

A. MAP OF M BED LEVEL

B. MAP OF K BED LEVEL (WORKINGS ON M BED LEVEL SUPERPOSED)



C. MAP OF G BED LEVEL (WORKINGS ON K BED LEVEL SUPERPOSED)



D. SECTION ALONG LINE *A-B*, TULSA-QUAPAW MINE, AND SECTION ALONG LINE *C-D*, BENDELARI MINE

Both sections show thinning of ore-bearing strata in dolomitized ground and adjacent strongly leached ground, superposition of dolomite cores on two levels, and restriction of M bed stopes to edges of dolomite. (See plate 5 for locations of Sections A-B and C-D)

100 0 100 200 FEET
DATUM IS MEAN SEA LEVEL

EXPLANATION

Limestone
Virtually unaltered, with nodular and bedded chert; chert may be brecciated

Partially leached limestone
Generally mottled to completely replaced by dolomite or jasperoid; residuum more residual clay and introduced shale; chert usually concentrated and brecciated

Strongly leached ground
Dominantly rough and porous, "boulder ground," in part, contains chert nodules and fragments with varying amounts of jasperoid, clay and shale, calcite, and limestone remnants

Gray spar dolomite matrix
Partly to completely dolomitized limestone. Dolomite is partly replaced by jasperoid. Ore-bearing where mined by stoping. Figures A, B, and C show only distribution of dolomite. Borders dashed where approximate; queries indicate mainly workings opened after completion of fieldwork

Jasperoid matrix
Partly to completely jasperoidized limestone or dolomite; forms thin films on chert nodules in strongly leached ground; ore-bearing where mined by stoping

Contact

Fault
Showing direction of movement

Workings on G bed level

Workings on K bed level

Workings on M bed level

Mine stope

Inclined workings between levels
Chevrons point down

Shaft

Winze

Property tie

See plate 5 for location of Bendelari and Tulsa-Quapaw mines within the Picher field

MAPS OF THREE LEVELS OF THE TULSA-QUAPAW MINE, AS OF 1940, AND DIAGRAMMATIC SECTIONS OF THE TULSA-QUAPAW AND BENDELARI MINES, KANSAS, ALL SHOWING RELATION OF STOPED GROUND TO THE GRAY SPAR DOLOMITE