Environment of Ore Deposition at the Mex-Tex Deposits, Hansonburg District, New Mexico, from Studies of Fluid Inclusions 1

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Abstract

These deposits, in Pennsylvanian limestone and shale, contain barite, fluorite, low-silver galena with “J-type” lead, and quartz, and only minor amounts of other minerals. Mineralization occurs in veins, in blankets of bedded, rhythmically banded “coontail” ore, and in vuggy, coarsely crystalline open-space fillings in tectonic and solution channels in limestone adjacent to faults. Except for widespread silicification, the mineralization is very similar to that of the southern Illinois deposits. The main stages of hypogene mineralization are: 1) sphalerite, pyrite, galena and chalcopyrite; 2) five easily recognizable substages of fluorite with intermittent quartz and barite; and 3) late calcite.

More than 500 primary and 2,500 pseudosecondary inclusions, mainly from fluorite, were studied on the freezing and heating microscope stages. Some primary inclusions contain organic matter. Gross supercooling indicates slow ore-fluid movement. Recognizable planes of pseudo-secondarys (each containing 20–200 inclusions) yield data essentially identical with coeval primaries, but the few planes of presumed secondaries do not.

The first three substages of fluorite formed from fluids that were essentially constant in temperature at 186°–205° C (assumed pressure correction +10° C), but increased in salinity from approximately 10 up to 15 weight percent salts. Succeeding substages formed at gradually decreasing temperatures (to about 140° C) and increasing salinity (maximum 17%), with breaks marking several individual substages. A few inclusions in an early barite have anomalously low homogenization temperatures. Coarse selenite has primary inclusions full of fresh water and hence is presumed to be supergene.

These data do not prove any given theory or origin for the fluids or the deposits, but they do place some limits on possible mechanisms of origin.

Geologic Setting of Ores

BARITE, fluorite, and galena deposits occur along faults in the Hansonburg mining district (Johnston, 1928) in southeastern Socorro County, N. Mex. (Fig. 1). The Mex-Tex group of mines is 28 miles east of San Antonio, New Mexico, on U. S. highway 380 and 5 miles from the Bingham post office, to the south of the highway. The main workings consist of a group of mines and prospects in sec. 36, T. 5 S., R. 5 E. and in secs. 30, 31, T. 5 S., R. 6 E. The Blanchard group lead mine (also called Portales or Hansonburg lead mine) is 1.5 miles south of the Mex-Tex group of workings.

All of these deposits lie along the westward facing fault-scarps of the Oscura Mountains in cliff-forming limestones, shales, and sandstones of the Magdalena Group of Pennsylvanian and Permian age (Kotlowski, 1953; Lasky, 1932, p. 65; Rothrock and others 1946). The sedimentary rocks rest on Precambrian granite in eastward tilted fault blocks which merge into a northward plunging anticline in the northern part of the mining district. Scattered sills and dikes of middle Tertiary quartz monzonite and diorite intrude the Pennsylvanian rocks. A relatively thin cover of Triassic, Cretaceous, and early Tertiary sedimentary rocks, which at one time overlaid the late Paleozoic rocks, is now eroded away. This cover probably did not exceed 1,500 meters.

The mineral deposits occur along northward-trending faults in the form of lenticular and blanketlike bedded replacement bodies and thin fissure veins (Fig. 2A). There is extensive replacement of the limestones by crystalline and jaspiloid quartz and by barite and fluorite. Rhythmically banded “coontail” ore, consisting of bands and tabular vugs of quartz, barite, and fluorite associated with galena and other sulfides, is in blankets that extend laterally outward from the faults in favored massive limestone beds (Fig. 2A). The blankets commonly occur on the eastward or updip side of the faults and extend along them for about one hundred meters.

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and hence are very similar to the manto-like bedded replacement deposits of "coontail" ore (Fig. 2B) in the southern Illinois fluor spar district (Weller, Grogan, and Tippie, 1952). In age, the deposits are at least post-Paleozoic, and pre late Tertiary.

The Hansonburg district is in the easternmost part of the Basin and Range physiographic province, just west of the high plains of the Great Plains province (Fenneman and Johnson, 1946). Thus the district lies along the margin of the central craton in which the alkaline suite of igneous rocks is most abundant (Brock and Heyl, 1961). Intrusive igneous rocks cropping out in the area are of calcalkaline, subalkaline, or alkaline affinities. They include quartz monzonite and diorite in the Oscura Mountains to the east and south; alkaline hypabyssal intrusions 30 miles eastward in the Carrizo Mountains in southwestern Lincoln County (Griswold, 1959; Perhac and Heinrich, 1964, p. 227), and the alkaline trachytes in the Gallinos Hills, 36 miles toward the northeast (Perhac and Heinrich, 1964). In addition, E. W. Heinrich states (written communication, 1966) that "alkaline rocks occur in the Tecolote Hills, Jicarillas

![Diagrammatic cross sections. A. Section facing northward across "upper" Mex-Tex ore body, Mex-Tex mine, Hansonburg district, New Mexico. Williams and others (1964, Fig. 17) show a very similar exposure, seen facing northward across northermost pit of the Royal Flush workings. Note the tabular blanketlike ore bodies on the up-dip side of the fault and the thin vein below, very similar to bedded ore bodies in southern Illinois as illustrated in Fig. 2B. B. Section of a bedded deposit in the Cave-In-Rock district of southern Illinois. Reproduced from Weller and others (1952, Fig. 18, lower section).](image)

[Mountains], Capitan Mountains, and Sierra Blanco," all between 30 and 40 miles eastward, and to the west of the Hansonburg district about 30 to 40 miles are "carbonatized (?) lamprophyres near Socorro," among otherwise nonalkaline mafic and felsic igneous rocks, such as the quartz monzonite near Magdalena.

**Mineralogy and Paragenesis**

Commonly the minerals of the deposits are large crystals that line fractures and vugs. Quartz, barite, fluorite, and galena are the main minerals. Fluorite and quartz were deposited during several periods, each with a characteristic color or crystal habit, in both the Mex-Tex group and the Blanchard mine (Fig. 3). Crystals of fluorite, galena, and quartz 5 cm across, and tabular crystals of barite as much as 15 cm long are common. Small quantities of sphalerite and pyrite were deposited early in the

2 Most of the early sulfides have been oxidized to many species of sparse, but beautifully crystalized supergene minerals. We are indebted to Dr. F. E. Kettlowski for the specimen of sphalerite, a deeply etched crystal 8 mm in diameter.
sequence of deposition, followed by minor chalcopyrite. The positions of minor amounts of tetrahedrite(?), siderite, and ankerite were not determined. Late calcite is probably the last hypogene mineral deposited. Several periods of leaching and replacement of the limestone wallrocks occurred during the general periods of hypogene mineral deposition, and at least two or three periods of faulting and renewed fracturing of the wallrocks near the faults.

The abundant silica in several successive stages is notable. The first stage consists of massive silification of the wallrocks, cementation of breccias, and selective replacement of limestone beds in the wallrocks. Bands of barite, galena and fluorite now fill the leached, unsilificated parts of the beds, and commonly have central, crystal-lined vugs. "Coontail" ore, consisting of a succession of such siliceous bands, forms the major part of the blanket ore bodies, similar to those in southern Illinois. Later quartz occurs in druzy crusts in the "coontail" ore and as large zoned crystals with amethystine or light smokey crystal caps. The most recent stage consists of thin druses of microscopic white quartz crystals coating the large ones.

The fluorite also reveals several stages of deposition, both by the superposition of isolated crystals and by color bands in zoned crystals. All the fluorite crystals are essentially cubic in habit, but occasional octahedral faces are found on crystals from several of the stages. The earliest stage (I) occurs as nearly round, pale green crystals, consisting of parallel, multiple-stepped cube facets, in part modified by the hexoctahedron, (124). These appear externally corroded, and have an internal zonal pattern that superficially appears to be the result of multiple leaching and deposition. However, all these zone boundaries are seen to be sharply regular crystal facets in detail, and hence are not typical of the smoothly curved leached surfaces frequently seen on other isometric minerals such as galena and sphalerite. Some of these crystals were then

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Fig. 3. Sequence of events during ore formation in the Hansonburg mining district, Socorro County, New Mexico.
covered with limpid single well-formed cubes of the same pale green color (fluorite II). Following this comes the cloudy, very slightly greenish white cubes of fluorite III and then limpid blue and blue-green cubes of fluorite IV. Finally, a very sparse lavender to purple fluorite, with blue bands, designated as stage V, formed mainly as a thin film over the earlier crystals.

The various stages of fluorite apparently were separated from each other by distinct breaks in the environment of deposition, possibly from changes in the chemistry of the solutions or in the "plumbing system." Individual crystals representing several stages have either abrupt color changes or a zone of small solid inclusions of quartz marking the break. Also, the growth of many crystals seems to have ended at a break in stage. As individual bands in the "coontail ore" contain zoned fluorite crystals from stages I through IV, we believe this "coontail" structure must have formed during or prior to stage I.

Some of the galena crystals have etched surfaces of hypogene origin. Although this might also signify a period of solution of the wallrock limestones, there are water solutions that could dissolve galena but not calcite. The galena of the district is relatively lean in silver, as a short ton of galena concentrates contains only a few ounces of silver. Silver-lean galena is typical of most lead deposits in the Mississippi Valley, but is uncommon in the western United States.

Previous Inclusion Studies in the General Area

Twenhofel (1947) measured the temperature of homogenization of inclusions in fluorite from the Greenleaf #1 prospect, Fluorite Ridge, Luna County, New Mexico (Russell, 1947), approximately 100 miles southwest of the Hansonsburg district. These inclusions had homogenization temperatures from a maximum of 202° C for an early, green octahedral stage to less than 150° for a later, pale violet, cubic stage. Ames (1958) studied the composition of the fluid in inclusions in galena, barite, and fluorite from the Blanchard group lead mine (Portales) in the Hansonsburg district and in fluorite from the Alamo mine, near Derry, New Mexico, approximately 80 miles southwest of the Hansonsburg district. He reports a very wide range in the relative amounts of K, Na, Cl, and SO₄ from inclusions in the four analyses, but the gross excess of anions over cations makes these results difficult to interpret. He also reports homogenization temperature determinations (by the U. S. Geological Survey) of 90° to 100° C for Blanchard fluorite. These are considerably lower than those found in the current work.

New Fluid Inclusion Studies

Sample Preparation and Examination Procedure. Doubly-polished plates, several millimeters thick, were cut through the centers of the larger crystals, to reveal the zonal arrangement. In addition, flat cleavage flakes from specific color zones were examined, using matching index liquid with cover glasses to improve visibility. Selected parts of these plates, containing inclusions of recognizable origin and relation to the zoning (Figs. 4–15), were then cut out for determination of the freezing and homogenization temperatures. Except as noted otherwise, all data and photographs are from samples from the Mex-Tex group.

The presence of presumably different composition fluids in the inclusions from the various zones in these samples was first revealed by differences in the behavior of their vapor bubbles in a thermal gradient (Roeder, 1966). Most of the bubbles were attracted by a warm probe, but some oscillated, and a few were repelled. Although the exact mechanism for such movement is not certain, the differences in behavior between the various groups of inclusions indicate, qualitatively, that there must also be differences in the composition of the liquid.

Origin of the inclusions studied

The fluorite samples had large numbers of excellent fluid inclusions, which could be identified as primary by several criteria as indicated in the figures. Many of these inclusions were large (>100 µ).

In addition to the primary inclusions many of the fluorites contained numerous almost flat planes of inclusions representing healed cleavage fractures. Although these frequently showed evidence of necking down of former larger inclusions, as seen in Figure 13, and hence might be expected to give spurious results, most of the inclusions in such planes that appear to contain no bubble are actually still connected through a very fine tube to an adjoining inclusion that does have a bubble. On heating, such planes of inclusions all homogenize within
FIG. 4. Primary and pseudosecondary inclusions in early fluorite I. Many thousands of tiny primary inclusions of dark organic matter and of brine outline a series of parallel and steplike growth stages at the bottom (A). Many of these inclusions started as flat imperfections parallel to the cube face that were covered over gradually, leaving behind a pointed conical cavity (see arrow: another similar but smaller inclusion, shown in Fig. 11, lies along one of the dark growth zones in this field of view, but is out of focus).

After deposition of a crust of quartz crystals (dark), additional fluorite growth occurred (B), practically free of inclusions and crystallographically parallel with the inner zone. A plane of probably pseudosecondary inclusions (PS) parallel to an octahedral cleavage direction crosses the earlier zone diagonally from the lower left and stops abruptly at the quartz layer. These inclusions have almost identical homogenization temperatures and freezing temperatures only slightly lower than the adjacent primary inclusions.
a very narrow temperature range. Most of these planes consist of pseudosecondary inclusions, trapped by the healing of fractures formed in the crystal during its growth (Figs. 12–15). The fluids they contain are thus the same as those trapped in primary inclusions that formed at the same time at the growing surface of the crystal. This interpretation was borne out by the fact that both the homogenization and freezing data on the pseudosecondary inclusions agree very closely with the data on associated primary inclusions. The process of fracturing and healing of the surface of the growing crystal has occurred many times, resulting in hundreds of such planes in some samples, each with hundreds of inclusions (Figs. 14, 15).

Freezing Temperature Determinations.—The freezing temperature of the inclusion fluid was determined on a microscope freezing stage (Roedder, 1962). For water solutions this freezing temperature is a function of the concentration of salts in solution. The precision and accuracy of such determinations will vary with the nature of the sample material. The maximum uncertainty here was ±0.5, and most were less than ±0.1°C.

Homogenization Temperature Determinations. After the freezing run, filling (homogenization) temperatures were determined with a Leitz heating stage, model 350, on many of the same inclusions used in the freezing studies, and on other similar inclusions. Rather large thermal gradients exist in this stage between the heating element, thermometer, and sample chamber during any change in temperature. The magnitude of these gradients, and their inevitable variation with change in sample size, placement, etc., precluded the use of any standardized “dynamic” procedure, such as the uniform rate of increase in temperature that is so frequently used in homogenization studies. A stepwise operation was used instead, with at least an approach to static conditions after each change. Calibration was achieved by the use of organic melting point standards, sealed in short sections of glass microcapillary tubes, mounted beside or between the sample plates. The compounds used were those melting at 136.0°, 149.5°, 174.5°, and 216.0°C, from a set of standard melting point reagents, unfortunately of unidentified composition, supplied for use with the Kofler heating stage (Arthur H. Thomas Co., Philadelphia). As a result of these calibration and operation procedures, most of the homogenization temperatures reported here have an uncertainty of ±5°C.

Verification of the absence of leakage during the homogenization determinations was obtained by the agreement between first and second determinations on some inclusions, and from bubble size before and after the run for all inclusions. Data from those few inclusions (<1 percent) that showed such evidence of leakage were discarded.

Data Obtained.—At room temperature, most of the inclusions consist of a colorless water solution and a small gas bubble, estimated at 5 to 7 volume percent, up to a maximum of nearly 10 percent (Fig. 7). A number of inclusions had a small (approximately 0.01 volume percent) unidentified birefringent crystal, possibly barite, that was not noticeably affected during homogenization runs. This might be a daughter mineral, formed from the inclusion fluid after trapping, or a solid inclusion, trapped by accident along with the fluid. If it is the former, its absence in many inclusions could be simply a result of failure to nucleate. Several of the primary fluid inclusions in fluorite are attached to crystals of quartz embedded in the host. These fluid inclusions were trapped when the host fluorite enclosed the solid crystal (Figs. 5 and 6).

Some of the inclusions showed the phenomenon of metastable “superheated” ice, at high negative pressures, due to failure to re nucleate the vapor bubble after its elimination on freezing (Roedder, 1967). Although these measurements of metastable equilibrium are consistent with the rest, all the data reported here are for stable equilibria.

The organic matter present in some inclusions was relatively unchanged at the temperature of homogenization of the aqueous liquid and vapor phases, and seemed to have no influence on the temperature; it has been ignored in reporting the temperatures.

Fig. 5. Primary inclusion trapped by imperfect enclosure of quartz crystal (q) in fluorite I (f). The fluorite grew upward as viewed here. This inclusion is present, but too small to be visible in the crust of quartz crystals of Fig. 4. Freezing temperature, −8.55°C; homogenization temperature, 185°C.

Fig. 6. Primary fluid inclusion in fluorite III. A number of very small unidentified grains and a few larger crystals of quartz (q) were deposited on the surface of a cube of inclusion-free fluorite (A). Further growth of the fluorite (B) trapped a primary fluid inclusion, which now contains liquid (l) and vapor (v).

Figs. 7 and 8. Clean, sharply faceted, isolated fluid inclusions within several millimeters of surface of late blugreen fluorite IV crystals. In each case these were almost the only inclusions for several millimeters in all directions in the crystal, and hence are almost certainly primary. A tiny unidentified birefringent crystal is just barely visible in Fig. 7 at this plane of focus (arrow). The freezing temperature of the inclusion in Fig. 7 was −9.85 ± 0.05°C, and the homogenization temperature was 166°C; the inclusions in Fig. 8 were lost during sample preparation and hence were not run.
Fig. 9. Primary inclusions of liquid and solid (?) organic matter outlining growth stages (left to right) in fluorite III crystal from vug in "coontail" ore. As each of the zones of inclusions starts abruptly at flat cube (and octahedron) planes, it is presumed that the organic matter was brought in with the ore fluids in discrete pulses and adhered to the fluorite surface. At higher magnification some of these inclusions are seen to consist of a yellow-brown liquid and vapor bubble; although the data are imprecise, these "oil" inclusions had homogenization temperatures approximately equal to those of adjacent primary inclusions of brine.

Fig. 10. Large primary inclusion that has trapped a mixture of brine and organic matter. The cloud of brownish granular organic matter contains one highly birefringent crystal (x) of unknown composition. The oval clear area (b) is a bridge of fluorite II across the center of what is, in effect, a ring-shaped inclusion. The vapor-liquid homogenization temperature (exclusive of organic matter) was 186°C.

Fig. 11. Inclusion trapped as a result of the fluorite I host (f) trying to cover over a flat imperfection of unknown origin (perhaps a spot of organic matter spread out on the surface—see arrow), during its growth from bottom to top. In the moderately saline brine (l) is a vapor bubble (v) and numerous dark brownish grains of unidentified organic matter. Freezing temperature, $-772^\circ \pm 0.08^\circ$ C; homogenization temperature (exclusive of organic matter), 180°C.
Some of the larger primary inclusions, particularly in fluorite, had spongy brownish masses of what is believed to be organic matter, which coalesced into smaller masses of darker brown, in part birefringent material during a cycle of freezing and thawing (Figs. 10 and 11). Several inclusions in fluorite IV which originally had no visible organic matter apparently did have some present, as indicated by their behavior during the homogenization runs. The vapor bubble shrank normally with temperature increase, to perhaps 0.1 volume percent, after which it did not disappear, even with 25° further increase. This tiny “bubble” was seen to have an index of refraction greater than the liquid, and hence is presumed to be liquid organic matter which coalesced on the surface of the vapor bubble during the run. In addition, inclusions of a brownish-yellow fluid, with small vapor bubbles, were found in the barite and fluorite III (Fig. 9). These were unaffected by freezing, and are assumed to be oil.

At room temperature the gas bubbles in the brine inclusions are under considerable pressure, as they expand 10- to 20-fold in volume when the inclusions are opened to atmospheric pressure on a crushing stage (adapted from that of Deicha, 1950). As this gas is very readily soluble in kerosene, it presumably consists of organic compounds such as methane and little if any CO₂.⁸

The inclusion salinity and temperature data obtained, summarized graphically in Figure 16, show one very notable and unique feature, an increase in the concentration of salts in solution from the earliest zone of the early fluorite (I) through to the core of the late quartz crystals. This is evidenced by the decrease in freezing temperatures from −6.7° to −13.0° C, corresponding to approximately 10.5 percent salts (NaCl equivalent) at the start of fluorite growth, to 17 percent during the growth of the quartz crystal cores. The concentration of salts in the fluids then dropped abruptly before the outer zone of the amethyst crystals grew, as four inclusions in this outer zone had freezing temperatures averaging −8.8° C. The same pattern of salinity increase is also evident in the secondary inclusions, in that most of them in any given zone had salinities comparable to the salinity of primary inclusions in later zones. The freezing data on pseudosecondary inclusions are uniformly comparable to those from adjacent or coeval primary inclusions.

Few determinations of freezing temperature have been made of inclusions from different generations or stages in a given paragenesis (Roedder, 1963; 1965). Most of these show a decrease in salinity (and usually also in homogenization temperature) toward the end of the mineralization. Hall and Friedman (1963, p. 902) report a small increase in salinity (a decrease in the freezing temperature from −17.8° for the center, to −19.0° for the edge of a single fluorite crystal), but their extensive analytical work indicates a pronounced and almost systematic decrease in concentration, for the various stages in both the Cave-In-Rock fluorite district in southern Illinois and the upper Mississippi Valley lead-zinc district. Indirect but valid evidence of similar decreases in salinity are found in the numerous reports of the presence of daughter crystals of NaCl and KCl in early primary inclusions that homogenize at high temperatures, and their absence in later primary or secondary inclusions that homogenize at lower temperatures, in specimens from both pegmatites and metal-bearing ore deposits (Ermakov, 1950).

The homogenization temperatures obtained are also unique, in that they are exceedingly uniform throughout the early stages. All 50 measured primary inclusions in the early fluorite I (Fig. 16) homogenized in the narrow range 177°–187° C. Twenty-four inclusions in the succeeding barite stage have much lower and variable homogenization temperatures. These are followed in turn by 176 primary inclusions in fluorite II and the inner part of fluorite III, which have once again a mode of 182° and a still rather narrow, but slightly greater spread (176°–192° C). After this essentially isothermal deposition, the temperature dropped 50° before the late quartz crystallized, with one apparent reversal at the start of crystallization of fluorite IV.

The sections cut from the barite from the Mex-Tex deposits had many inclusions, but only a few that seemed to be primary and runnable. These inclusions had considerably lower homogenization temperatures, with a 74° range. There are several possible explanations for this abrupt change. First, there is evidence of recrystallization and necking down in many of the barite inclusions, and so these inclusions may also have been affected, even though carefully selected. Second, it should be noted that the salinity and homogenization temperature of these inclusions correspond best with those of later fluids, so it is possible that they are actually secondary inclusions, in which gross recrystallization has eliminated evidence of secondary origin. Third, there is abundant evidence of leakage in the inclusions in many barite samples. It is thus possible that fluid leaked into these inclusions at some later period. Fourth it is possible, as discussed below, that the

⁸Barton (1967), in a discussion of the possible role of organic matter in the precipitation of Mississippi Valley ores, shows that H₂, O₂, SO₃, H₂S, and CO₂ all have too low pressures to contribute significantly to this gas expansion.
Fig. 12. Plane of pseudosecondary inclusions in fluorite. Section cut parallel to (110) and here viewed perpendicular to (110), with the plane of focus within the plate. The fluorite IV crystal grew upward as viewed here. The surface of the crystal is at (B), and has a reentrant at the left that might have become a large inclusion had further growth occurred. A plane of inclusions parallel to an octahedral cleavage fracture ends abruptly at (A), well within the solid crystal, proving that these are pseudosecondary inclusions; that is, the crystal grew to (A), fractured and healed, trapping inclusions, and then grew further. Any plane of inclusions that is intersected by the plane of the section will appear this way, but in such a case the edge of the inclusion plane similar to (A) will be at the surface and not be actually embedded as this is. Note that the fracture tapered down toward the bottom, as indicated by the decrease in inclusion.
temperature drop at the barite stage is real. We cannot exclude any of these possibilities at this time.

The pressure correction to add to the homogenization temperatures of Figure 16 can only be guessed, as we do not know the depth of cover at the time of mineralization. Assuming this cover to be at its maximum probable value of 1500 meters, and using the data of Klevtsov and Lemmelin (1959) for 10% NaCl, this correction would fall between +10° (hydrostatic load) and +30° (lithostatic load).

Although the data plotted on Figure 16 for primary inclusions in any given stage show gross overlaps with adjacent stages in the temperature ranges particularly for homogenization, most of these overlaps are the result of individual sample variation. Thus when the various zones in a given specimen are compared (as indicated by lines at the top of the figure), the changes with stage become much more systematic, and the degree of overlap becomes reduced. Furthermore, the data for most (but not all) of the individual blocks plotted on Figure 16 actually have similar but internal, uncharted trends. Similar effects, though less noticeable, were found in the freezing data, but these seem to indicate significant reversals at the start of stage II and stage IV fluorite. It thus appears that not only were local differences superimposed on the major changes in salinity and temperature, but that at least some of the apparent breaks in the paragenetic sequence, indicated by changes in color or habit, represent real and abrupt changes in the ore fluids.

Discussion and Speculation

We can never know what the environment of ore deposition was at the time these deposits were formed, but the inclusion evidence permits us to place rather specific limits on some of the possible variables. The fluids from which these ores formed were moderately to strongly saline brines, probably high in chloride, with temperatures near 200° C except for the very last stages. Suspended organic matter was also present at times, probably mobilized from sediments. These fluids, rising along the Oscura fault to the shale of the Permian Bursum Formation and then spreading laterally through the Paleozoic limestones, were probably undersaturated with respect to calcite, except in the final stage, as evidenced by the extensive solution cavities in the host limestone, and also by the "coontail ore" (under the assumption that this is formed by some process of repeated crustification and solution collapse). They were saturated, at various times, with respect to galena, barite, fluorite, and quartz. They were probably moving slowly and were close to thermal equilibrium with the country rock, as large, well-formed clear crystals normally do not form under the fluctuating and unstable physical and chemical conditions expected in most natural fast-flowing fluids. Slow movement is also indicated by the surprising uniformity of homogenization temperatures throughout the early stages of mineralization, and by the extensive supercooling of the fluid inclusions observed during the freezing studies (Roedder, 1962, p. 1053).

The sudden drop in temperature at the barite stage is difficult to explain, in view of the uniformity of temperature before and after. If an ore fluid of uniform temperature flowed through a fissure in cooler rock, on its way to the present site of the ore body, the walls would eventually have come to essentially the same temperature as the fluid. If a new bypass for the fluids through still cool rock then opened up along the route, the fluids reaching the site of the ore body would be cooler, until wallrock along this new route also has been brought up to the original fluid temperature. Such an event would possibly cause a change in the nature of the solids inclusions, the majority of which are too small to be resolved at this magnification. Photograph was taken with well-collimated lighting; in ordinary light this plate merely looks cloudy. Figure 13 is a photograph of similar but much larger fractures in another sample.

Fig. 15. Planes of pseudosecondary inclusions in fluorite V. Section through the edge of a 4-cm cube cut parallel to (110), showing several sets of (111) cleavage fractures, now outlined by pseudosecondary inclusions. The surface of the crystal (now at B) was fractured repeatedly during its growth. The main part of these fractures occur in pairs of nonadjacent octahedral cleavage surfaces, such as (111) and (111), with the line of intersection at or near the surface of the crystal at the time of fracturing. At (A) the line of intersection of the pair is in the plane of the photograph; at (C) it is perpendicular to the plane of the photograph. In each plane the inclusions are graduated downward in size toward the center of the crystal, indicating that the crack was widest at the surface of the crystal and tapered inward. Three large primary inclusions (P) are also visible. Sample from the Hanzenburg lead mine.
Fig. 16. Freezing and homogenization temperatures on inclusions, plotted against paragenetic sequence. A pressure correction of 10° to 30° C must be added to the homogenization temperatures to obtain the formation (trapping) temperatures. The following are not shown: 1), data on a small number (usually <2 percent) of the inclusions in some planes of secondaries that had metastable "superheated" ice (Roedder, 1967), 2), data on a smaller number (<1 percent) that leaked visibly; 3), one plane of about 30 apparently very late secondary inclusions in the outer shell of a fluorite I crystal, with freezing temperatures between −0.70° and −1.80° C; and 4), data on a large number of primary inclusions in a colorless, glassy selenite crystal, 20 cm in diameter, which contained essentially fresh water and no bubble (these, and the selenite, are thus assumed to be supergene). Dots indicate individual primary inclusions; for ease of plotting, groups believed to be cogenetic are indicated by boxes showing the range of values and mode for a specific or estimated number (30 or more) of inclusions. These occurred in groups of definite primary origin (P); or in planes of pseudosecondary (PS) or secondary (S) origin. For most samples both determinations were made on the same group of inclusions. Where several different planes were involved, identifying letters are added. Where different zones in the same crystal were studied, the pertinent columns of data have been connected by lines at the top of the diagram. The sphalerite and fluorite V samples are from the Hansonburg lead mine; all others are from the Mex-Tex group.
precipitated (as occurred here), but would not be expected to alter the salinity of the fluids greatly. Although there is evidence of several periods of faulting during the formation of the ore body (Fig. 3), none of these coincide with the start of formation of barite. Faulting does coincide with much of the later part of the barite stage. This does not preclude the suggested mechanism, as the movements to form such a new bypass could well have been at a considerable distance from the present ore body and not affected it.

The combination of overall uniform homogenization temperatures together with a considerable increase in salinity is somewhat paradoxical, as mineralization almost certainly occurred under conditions of greater-than-normal geothermal gradient. Mixing of near-surface meteoric water with deeper brines is precluded by the combination of uniform temperature with an increase in salinity (except perhaps during the very last amethyst quartz stage). Slawson and Austin (1962) showed that the Hansonburg district fits into a regional pattern of lead isotope variations that can be explained by assuming a rise of homogeneous solutions from a deep source, along a very large structural linament, to yield a number of small deposits. A deep source could provide an essentially isothermal reservoir of fluids, but would not readily yield the change in salinity.

One possible interpretation is that the rise in salinity with paragenetic stage is due to the flushing out of earlier fluids by new, more saline fluids introduced from below. This interpretation is in accord with the evidence presented by Austin and Slawson (1961), from a study of lead isotopes in zoned galena crystals from the nearby Blanchard group (Hansonburg) mine. This showed a zoning pattern which could be explained by the movement of an originally uniform, homogeneous ore fluid, the first portions of which flushed out appreciable amounts of radiogenic lead from the rocks traversed, yielding crystals with more radiogenic lead in the cores. Unfortunately, all of the fluorite stages sampled in the present study were deposited after the end of the galena mineralization. Furthermore, Figure 16 does not include data on the large volumes of early fluids that must have circulated to cause the gross solution and silicification of limestone, as well as deposition of most of the early sulfides. Unless there were major changes in the plumbing system leading to the deposit after this stage, these early fluids would probably have flushed out all preexisting solutions along the route.

It is unfortunate that only one small crystal of sphalerite was available for study, the rest having been removed by surface oxidation. The inclusions in this sphalerite had lower homogenization temperatures and slightly higher salinities than the succeeding fluorite I. Only two of these were recognizably primary, and the sample came from the Hansonburg mine, rather than the Mex-Tex, so any evidence of overall trends is dubious, but the data do suggest a general rise in temperature to a maximum at fluorite II. It would obviously be desirable to examine inclusions in the early silification quartz, but this material is too fine grained.

Another possibility that might be considered (P. M. Bethke, oral communication, 1967) is that the sequence of fluids of increasing salinity represents progressive dewatering by compaction of shales in a sedimentary basin. Ion filtration would cause a progressive increase in salinity of the emerging waters with time and could be relatively isothermal.

The deposits are sufficiently similar to those in the Mississippi Valley to suggest a genetic similarity. As shown above, the Hansonburg district deposits are structurally very similar to the Illinois-Kentucky "coontail" bedded deposits of the Cave-in-Rock district, Illinois. The ores in both districts are relatively simple mineralogically, are notably coarsely crystallized, and have similar repeated bands of crustiform minerals. In both districts the galena is relatively lean in silver and has notably "J-type" lead isotope ratios. Both districts show evidence of solid, liquid, and gaseous organic matter in the trapped fluids, and both have similar solution collapse structures in the replaced limestones beneath capping shales (Grogan, 1949). Alkaline igneous rocks are reported in or near both districts, but basaltic and granitic igneous rocks are also common in the part of New Mexico in which the Hansonburg district is located.

Fluid inclusions in the two districts are not so similar. The Hansonburg samples have slightly higher homogenization temperatures, and considerably more dilute brines than those in the Illinois-Kentucky deposits. The Hansonburg ore fluids also increased markedly in salinity during ore deposition; in contrast, those in southern Illinois became progressively more dilute (Hall and Friedman, 1963). An even more marked progressive dilution of fluids was noted in the Wisconsin ores, but this district is free of fluorite. The Hansonburg inclusions have considerably higher salinities, however, than inclusions in most of the vein-type lead-zinc-silver-copper deposits of the Basin and Range province, as at Creede, Colorado (Roedder, 1963), and in this respect they are more similar to the Cave-in-Rock district.

Thus the inclusions indicate that the Hansonburg ores were deposited from an ore-bearing fluid that may be of a transitional type. Hansonburg lies between the little-disturbed alkaline igneous rock pro-
province of the craton of the central United States to the east and the normal igneous rock province of the Basin and Range province to the west, in which alka-
lic rocks are quite uncommon. If some fraction of
the fluids that deposited these ores is of deep origin
one might thus expect that this would also be transi-
tional in nature.

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