Mississippi Valley-type lead-fluorite-barite deposits of the Hansonburg mining district, New Mexico

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MISSISSIPPI VALLEY-TYPE LEAD-FLUORITE-BARITE DEPOSITS OF THE HANSONBURG MINING DISTRICT

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INTRODUCTION

Genetic models for Mississippi Valley-type deposits have long been discussed. A wide variety of models have been hotly debated (Brown, 1967); examples are the early-diagenetic model of Davis (1977) and ideas postulating ore deposition from two fluids (Ohle, 1980).The pioneering work by White (1958) and others suggests the active involvement of basinal, connate brines in the transport of metals. An enigma concerning Mississippi Valley-type deposits is the size of the districts and the uniform style of mineralization within the district. In this, the Hansonburg district is no different. It is 5 km long with virtually identical mineralization over the entire district. Similar mineralization (galena, fluorite, and barite) occurs sporadically along the entire Rio Grande rift in New Mexico.

The mineralization of the Hansonburg deposit was studied in detail in order to obtain sufficient data to form a coherent genetic model for a Mississippi Valley-type deposit. The study consisted of microthermometry and microanalysis of fluid inclusions from minerals in paragenetic sequence from several deposits in the Hansonburg district. Thermodynamic calculations based upon the analyses yielded pH, f_{oe}, and f_s, of the mineralizing solutions. These, coupled with field observations, were used in modeling the genesis of the deposit.

PREVIOUS WORK

First mention of the Hansonburg deposit was by Lindgren and others, in 1910. Lasky (1932) and Kottlowski (1953) published the first detailed reports on the district and noted its similarity to the Mississippi Valley deposits of the central United States. Kopicki (1962) and Lewchalermvong (1973) discussed the general geology and genesis of the deposits.

Slawson and Austin (1960) published lead-isotope studies on galena from a number of New Mexico deposits, and indicated that the Hansonburg district has highly radiogenic J-type leads. Roedder and others (1968) delineated a complex fluorite paragenesis based upon fluid inclusion thermometry studies, and classified the Hansonburg as a Mississippi Valley-type deposit. Beane (1974) suggested that the source of the lead was the overlying Permian arkosic units, which in turn were derived from Precambrian granitic rocks. Allmandinger (1974) concluded from stable-isotope studies that Permian evaporites supplied sulfur to the ore solutions. Ewing (1979) suggested that leaching of Precambrian-aged basement rocks supplied radiogerqc leads to the fluids.

GEOLOGIC SETTING

Structure

The Hansonburg mining district is located in the Oscura Mountains, on the eastern margin of the Rio Grande rift (fig. 1). The area is characterized by north-trending, en echelon normal faults which resulted from regional extension related to the Rio Grande rift. Southeast-trending, high-angle faults which cross-cut the Oscura Mountains are prominent in the Hansonburg district (fig. 2). These faults show little vertical offset, and in the vicinity of the Mex-Tex deposit show a reverse-normal displacement. The attitude of these faults, their restricted occurrence within the Hansonburg region, and their reverse-normal nature suggest that they did not result from rift-related extension but rather shearing (right-lateral?) which parallels the Capitan lineament.

The Capitan lineament (Kelley and Thompson, 1964; Chapin and others, 1978) passes through the Hansonburg region (fig. 1). It trends west-northwest and is defined in part by the alignment of Tertiary intrusives, including the Jones Camp dike, the basaltic craters of the Carrizozo malpais, the Capitan intrusive and the lengthy dikes east of Roswell, New Mexico (Chapin, 1979).
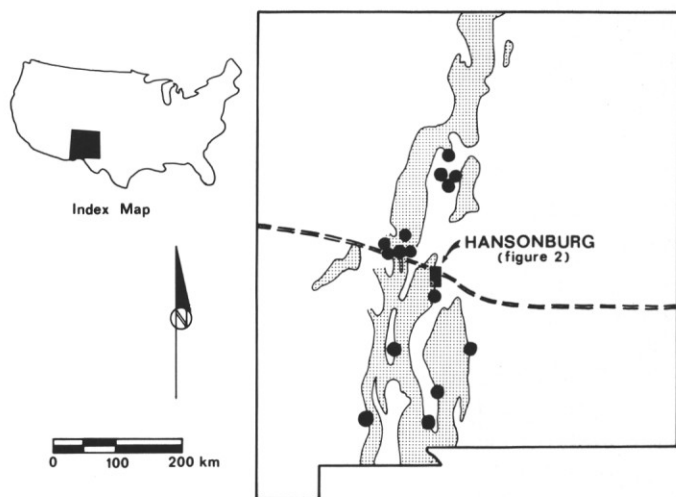

Figure 1. Location map showing location of the Rio Grande rift (stipple pattern), Capitan lineament (dashed line), Hansonburg district, and lead-fluorite-barite deposits (black dots) (modified after Beane, 1974).
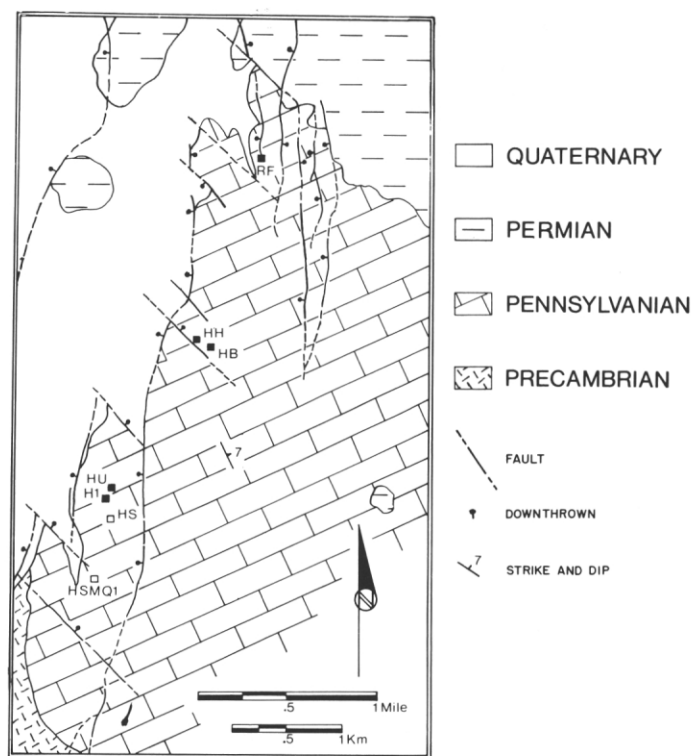
Figure 2. Generalized geologic map of the Hansonburg district showing sample locations by open (surface) and solid (underground) squares.

Figure 3. Outcrop of Council Springs Limestone showing bedding-plane anastomosis developed in more porous beds.

The Capitan lineament most probably is the east-west paleostructure referred to by Eardley (1962), and again by Kelley and Thompson (1964), along which there were repeated uplifts in south-central New Mexico during Paleozoic and Mesozoic periods. These uplifts affected the distribution, thickness, and lithologic character of Paleozoic and Mesozoic sediments (Wilpolt and Wanek, 1951) and indicate prolonged structural activity along the lineament. Thinning of stratigraphic units, such as the Paleozoic section over the Capitan lineament, is typical of other lineaments elsewhere in the western United States (Weimer, 1978).

Stratigraphy

Precambrian crystalline granitic rocks form the basement in the Hansonburg district. This basement is overlain by Pennsylvanian and Permian formations consisting of limestones, sandstones, arkoses, and evaporites. These Paleozoic sedimentary rocks are intruded by dikes and sills of Cenozoic age (Wilpolt and Wanek, 1951).

Paleozoic rocks in the vicinity of the Hansonburg district consist of marine limestones, shales, and sandstones of Pennsylvanian age. These units are thinly to massively bedded, and contain abundant fossil and organic debris. The Paleozoic section thins in the Hansonburg area suggesting a paleotopographic high across which deposition was interrupted and/or erosional truncation occurred. Stratigraphic correlations in southern New Mexico support the existence of a paleohighland in the Hansonburg region throughout Pennsylvanian time (Kottlowski, 1960, Pls. 2, 3). It is plausible that movement on the Capitan lineament resulted in the thinning of the Paleozoic section in the Hansonburg area and also localized the reef-facies limestone which hosts the ore deposits.

The mineralization in the Hansonburg district occurs in the Council Springs Member of the Pennsylvanian Madera Limestone. The Council Springs is predominantly a very fine grained, massive, silicified limestone locally containing lense-shaped cavernous zones, which range from small-scale, bedding-plane cavities (fig. 3) to room-sized caves. This secondary porosity is best observed in areas devoid of minerali

zation. The Council Springs possesses an erosional upper contact, characterized by pot-holes (Kopicki, 1962) and other karst-like features.

In areas of solution cavities, the Council Springs member grades from massive to a banded, reef-facies limestone. The reef-facies limestone consists of alternating bands of very fine micritic detritus and interbeds of coarse allochthonous fragmental carbonate with minor pyrite (fig. 4). Most of the coarse fragments are quite porous and appear to have been originally composed of carbonate algal material; however, dolomitization and silicification of the limestone makes identification of these lithic fragments difficult. The porous nature of the fragments greatly increases the primary porosity of the limestone unit. The bedding-plane cavities, which commonly are mineralized, resulted from dissolution of the coarse fragmental layers (fig. 4). Continued dissolution resulted in caves filled with rubble composed of limestone blocks with bedding-plane cavities.

The time of solution of the Council Springs Member is not clear. However, the occurrence of an erosional upper contact on the Council Springs Member with karst-like features indicates that solution may have occurred during Pennsylvanian time.

ORE DEPOSITS

Mineralization

Mineralization in the Hansonburg district is exposed over a 5-km distance along the normal-faulted, north-trending, eastern margin of
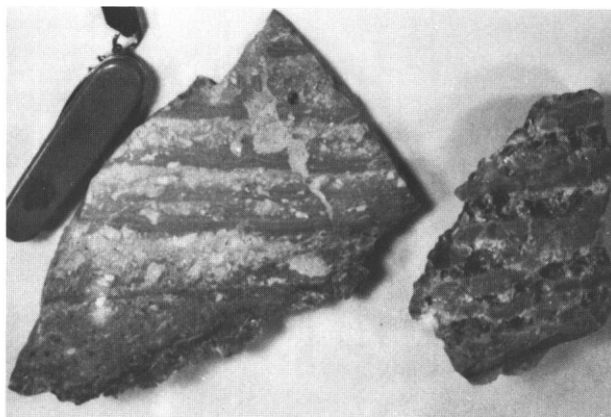

Figure 4. Examples of unmineralized (left) and mineralized (right) allochthonous limestone. Note the excellent correlation between the coarse, fragmental banding and the localization of the purple fluorite mineralization (dark bands). Fine-grained, micritic interbeds are left relatively unaffected by the karstification and later mineralization.

the Oscura Mountains (fig. 2). Known occurrences of mineralization are exposed over an east-west distance of 0.5 km in numerous blocks cut by normal faults. Reports of exploratory drilling indicate mineralization has been intersected both to the west and to the east of the exposed mineralization (R. S. Steensma, 1980, oral commun.). There is no evidence that the normal faults acted as conduits for the mineralizing solutions.

The main-stage hydrothermal mineralization in the Hansonburg district consists of galena, barite, fluorite, and quartz with minor sphalerite, pyrite, and chalcopyrite. The principal ore occurs as discontinuous pods and vugs occupying solution ways within the karstified Council Springs unit. Replacement of the limestone is minor and consists of galena and barite. Alteration consists of silicification and dolomitization of carbonate. The mineralization exhibits a layering which reflects the paragenesis. This has been described as a banded, or a "coontail" ore (Kottlowski, 1953; Roedder and others, 1968) after similarly named ores in other Mississippi Valley-type deposits.

Siderite occurs as fracture fillings in carbonate units above the mineralization and as fillings in cavities devoid of other minerals. The relationship of siderite to other mineralization in the Hansonburg district is unclear. Anderson (1957) mentions the occurrence of tennantite with a dolomite gangue in areas devoid of lead mineralization. We have not observed any of this type of mineralization.

Paragenesis

The general paragenesis in the district is early silicification and dolomitization with minor replacement of limestone by barite and galena. Locally, near the southeast-trending faults, cavities were first filled with terminated quartz crystals. This was followed by main-stage, open-space filling by galena with minor pyrite and spalerite followed by barite, fluorite, and quartz. Locally, all phases coprecipitated. Chalcopyrite was a late-forming phase and occurs principally with pyrite as minute inclusions within fluorite. Massive selenite fills vugs in mineralized areas but its source was most probably the overlying Permian anhydrite beds.

The paragenesis here stated is somewhat different and more variable than that proposed by Roedder and others (1968). This is due, in part, to the greater number of mineral occurrences studied during our investigation. Fluorite coloration shows quite wide variability in the many mineralized sites; thus, attempts to apply Roedder and others' fluorite-coloration paragenesis to the district proved fruitless.

FLUID INCLUSION STUDY

Samples were obtained from seven occurrences of mineralization (fig. 2). Quartz from both early and later mineralizing stages, fluorites of varying color and age, and barite were sampled at each locality. Sphalerite was sampled at only one locality. More than 1200 inclusions from 40 doubly polished thick sections were studied. Primary, pseudosecondary, and secondary inclusions were observed. Primary inclusions are quite large, generally greater than 100 microns, and abundant in all minerals examined. Most inclusions are simple two-phase types (liquid and vapor), with the vapor bubble occupying 1-5 percent of the inclusion volume. An immiscible hydrocarbon phase was observed in some inclusions as was also noted by Roedder and others (1968). A few inclusions appeared to contain mostly hydrocarbons.

Microthermometry

Homogenization temperatures ranged from 125-210°C and salinities from 10-18 equivalent wt.% NaCl (fig. 5). Results are similar to those reported by Roedder and others (1968) for the Mex-Tex deposit. There are, however, significant differences between the data from different deposits.
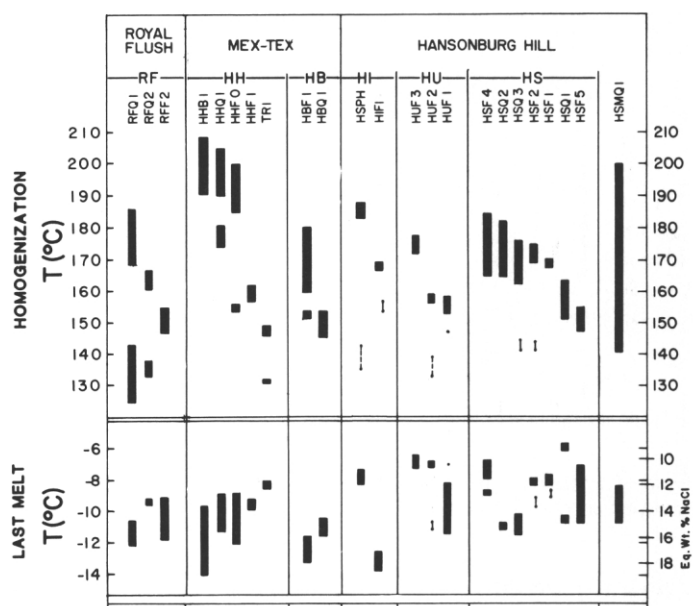

Figure 5. Fluid-inclusion homogenization and freezing data plotted against sample location, north to south (left to right). Data for each sample location (RF, HH, HB, etc.) are plotted in paragenetic sequence, with data from the earliest mineral phase observed plotted on the left of each column. HSMQI represents a large, multistage quartz crystal (see text and fig. 6). A pressure correction of 10–30°C must be added for true temperature of mineralization (Roedder and others, 1968). Dark bars indicate ranges of observed temperatures. Gaps in dark bars indicate groups of data from different growth periods within the same crystal. Higher temperatures were observed at the roots of crystals (early growth), with later growth-zones showing cooler temperatures.

Higher homogenization temperatures appear to be related to southeast-trending structures. The highest filling temperatures were measured in the Mex-Tex and HSMQ samples. The location from which these samples were collected is much nearer southeast-trending faults than the other 4 occurrences sampled, which have lower overall fluid-inclusion filling temperatures.

The data from different sample locations (columns RF to HS, fig. 5) all indicate the mineralizing solutions were cooling during the course of mineralization. An overall district-wide decrease in temperatures of the ore fluids is indicated from north to south.

Changes in salinity exist through the mineral paragenesis in each of the locations sampled. Salinity decreased slightly with time at the Royal Flush and Mex-Tex occurrences; salinity appears to have increased slightly with time at all four sites sampled on Hansonburg Hill. A district-wide decrease in fluid inclusion salinities is apparent from north to south (fig. 5).

A large quartz crystal (sample HSMQ1) from the extreme southern end of the district showed remarkable temperature variations through the growth of individual crystal zones, which may represent a complete record of the temperature variations of the mineralizing solutions responsible for the Hansonburg deposits (fig. 6). This quartz came from an occurrence along one of the southeast-trending structures, and is the only mineralization present at that site. The sample indicates that early, hot, mineralizing solutions entered the Council Springs unit and precipitated quartz along cavern walls during a period of wildly-fluctuating solution-temperatures (200-142°C, Root Zone-A) but with a fairly uniform salinity of 12-14 eq. wt.% NaCl. Through Band C, the fluctuations in temperature diminished and the temperature of the fluids appears to have decreased with time. The rise in temperature and increase in the range of filling temperatures (184-158°C) in zone D suggest a second
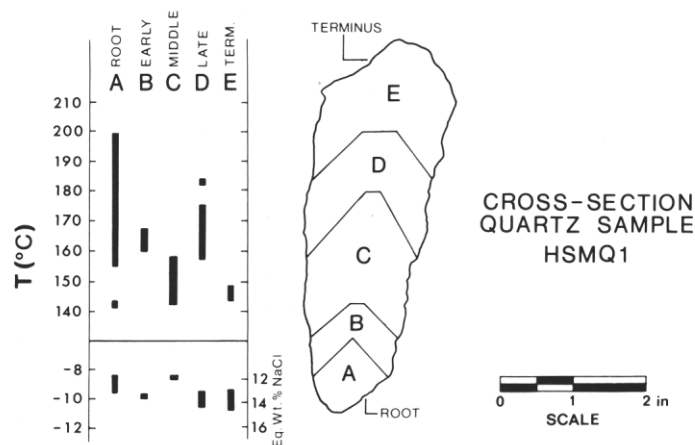
Figure 6. Fluid inclusions homogenization data plotted against crystal growth zones, A–E (see text).

	Galena (average of 5 samples)	Fluorite (average of 9 samples)	Late Quartz (average of 4 samples)
Analysis			
Ca(ppm)	8,400	*	3,300
Mg(ppm)	80	580	130
K (ppm)	5,800	2,000	1,400
Na(ppm)	51,000	45,000	39,000
Zn(ppm)	600	< 50	900
Cu(ppm)	< 10	200	1,000
Pb(ppm)	*	380	1,500
Fe(ppm)	110	600	1,100
Cl(ppm)	96,000	24,000	48,000
F (ppm)	930	*	400
CO_2(wt.%)	0.87	1.4	1.1
Temperature (°C)			
Avg. Calc.(Na/K)[1]	218	164	148
Avg. Meas. (T_h)	–	160	162
Salinity Calc. from Analysis (eq.wt.% NaCl)	15.5%	7.3%[2]	9.0%

episode of mineralization. The terminus zone of the crystal (zone E) has lower filling temperatures with little variation suggesting the mineralizing system was dying out.

Roedder and others (1968) estimated the maximum pressure corrections for fluid-inclusion homogenization temperatures to be 10 to 30°C for the Mex-Tex deposit. We agree with their estimate.

Composition of Fluid Inclusions

Inclusion fluids were extracted from galena, fluorite, and quartz and analyzed by methods similar to those described by Roedder and others (1963), and by Rye and Haffty (1969); see Putnam (1980) for details. Inclusion liquids from barite were not analyzed because barite had few primary fluid inclusions.

Fluid-inclusion waters in minerals from the Hansonburg district are Na-Ca-K-CI brines carrying hundreds of parts per million metals in solution (Table 1). Agreement exists between the measured salinities and those measured by freezing-point depressions. Freezing-point depressions are several degrees lower than predicted from the analyses, but this can be attributed to hydrocarbons and high amounts of CO, and N, which we have measured in the inclusion fluids.

Cation-to-anion balances indicated insufficient anions relative to the amounts of detected cations. This was also observed in analyses of inclusion fluids in the Cave-in-Rock and Upper Mississippi Valley districts (Hall and Friedman, 1963). The reason for this imbalance is not clear.

Na/K temperatures were calculated using Norman's (1977) modification of Fournier and Truesdell's Na-K-Ca geothermometer. Good agreement exists between actual measured fluid-inclusion homogenization temperatures and the calculated temperatures from Na/K ratios in inclusion fluids from fluorite and quartz (Table 1). The average calculated temperature for galena deposition was 218°C which is in agreement both with galena being an early crystallizing phase and with measured fluid-inclusion-homogenization temperatures that indicate temperatures near 200°C early in the paragenesis.

Thermodynamic Calculations

The pH of the ore solutions was calculated to be 4.5 at both 150 and 200°C using the measured CO, and calcium molalities, and assuming equilibrium between the ore fluids and calcite. This assumption is consistent with field observations which indicate a predominance of open-space filling, with only minor replacement of the limestone.

The maximum oxygen fugacity (log f_{o2}) during mineralization was calculated to be − 43 at 150°C and − 39 at 200°C, assuming equilibrium

between the mineral pairs galena-anhydrite, sphalerite-anhydrite, and pyrite-anhydrite, together with the analyses from Table 1. The f_s was calculated assuming similar equilibrium relations, and indicates a log f_s for the ore fluids of − 18.5 to −16 at 150°C and − 16 to − 13.5 at 200°C. The log activity of H_2S was calculated to be − 3.6 at 200°C and −7 at 150°C using measured Ph" and Cl⁻ concentrations and assuming $PbCl_4$= complexing.

Oxygen fugacity-pH diagrams were constructed (Putnam, 1980) and show the region (stipled) defined by the analyses of Hansonburg minerals (fig. 7). Results of mass-transfer calculations (PATH, Helgeson and others, 1970) are shown as a diamond on the diagrams.

The diagrams indicate that later solutions (150°C) favor the formation of sulfate rather than sulfide minerals. This agrees well with the notable lack of late-stage sulfide minerals at the Hansonburg deposit, whereas barite ($BaSO_4$) is common. Oxygen-sulfur fugacity diagrams were also
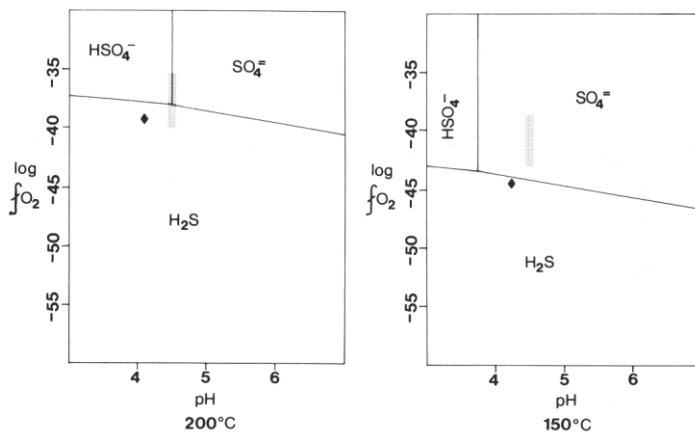

Figure 7. Stability fields for sulfur species at 200°C (left) and 150°C (right). Stippled region is defined by analysis of Hansonburg fluids. Diamond indicates PATH (Helgeson and others, 1970) calculations for Hansonburg fluids. Calculations are based upon averaged fluid compositions from Table 1 and equilibrium constants in Putnam (1980).
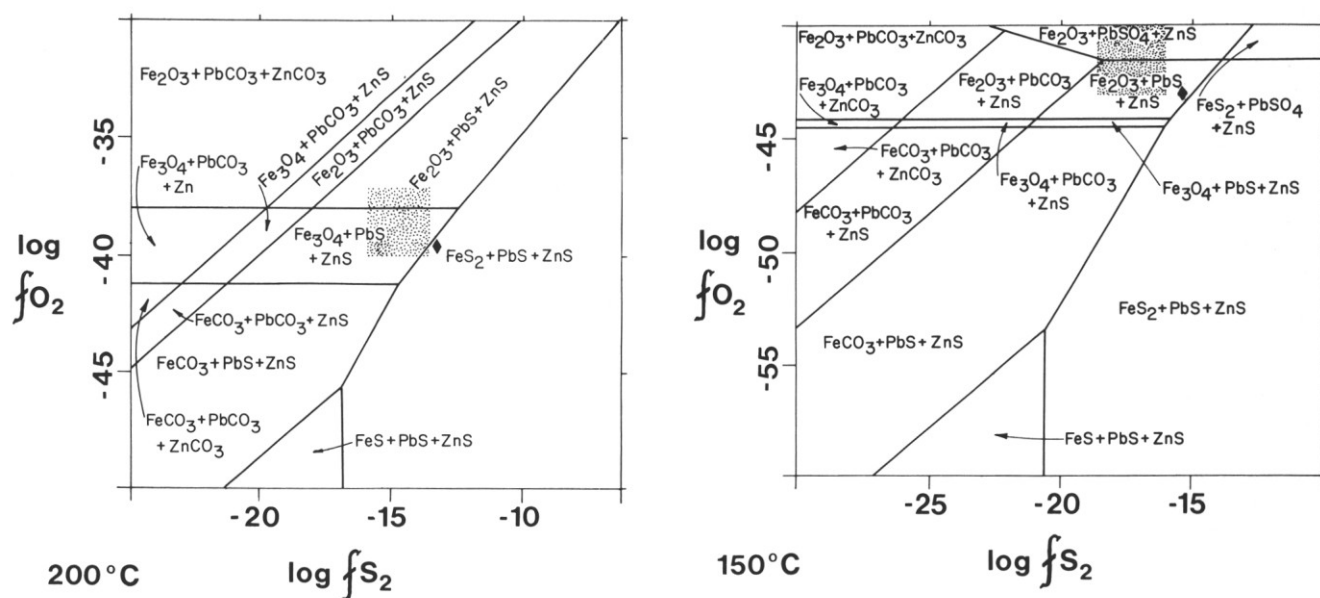
Figure 8. Stability fields for indicated mineral phases at $f_{CO_2}=39$ at $T=200°C$ and $f_{CO_2}=44$ at $T=150°C$. Stippled region is defined by analysis of Hansonburg fluids. Diamond indicates PATH (Helgeson and others, 1970) results for Hansonburg fluids. Calculations are based upon averaged fluid compositions from Table 1 and equilibrium constants in Putnam (1980).

constructed (Putnam, 1980) and indicate that early solutions (fig. 8) were predominantly in equilibrium with sulfide minerals (galena, sphalerite, and pyrite) while later, cooler solutions (150°C) were in equilibrium with sulfates, carbonates, and minor sulfides. These results again agree well with field relations. Overall the calculations indicate a relative increase in oxygen fugacity as the temperature decreased.

DISCUSSION

Pressure and Depth

Pressure at the time of mineralization had to be greater than 70 bars based on the combined $f_{..}$, and $f_{ir,o}$ of the ore fluids. This implies a depth beneath the water table of greater than 700 m at the time of mineralization in a hydrostatically pressured system (Haas, 1971).

Geochemistry

The geochemical calculations show that at 200°C solutions with hundreds of parts per million metals could also carry reduced sulfur with log $a_{,,s}= -3.6$ (see also fig. 7). Measurements of fluid inclusion gases in minerals from the Hansonburg deposit, which will be reported elsewhere, indicate ore fluids have approximately 50 ppm H,S. Mineralization can simply be attributed to cooling of the ore fluids and/or an increase in pH due to minor CO, effervescence. The change from sulfide to barite deposition through time can be explained by the increased relative f_{o2} of the fluids as the temperature decreased. This resulted in the H,S/SO, ratio in the ore fluid decreasing approximately by a factor of 5000 and hence a change from dominantly sulfide to dominantly sulfate mineralization.

Most hydrothermal deposits are thought to be the result of deposition by cooling or chemical change of an ore fluid carrying all the mineralizing components. However, a one-ore-solution theory for Mississippi Valley-type deposits has many opponents (Barnes, 1979). One of the factors these opponents have perhaps failed to consider is that the solutions may have a low pH (hence can carry significant concentrations of metals and H,S) with high amounts of CO, (hence will not attack the carbonate host). Such conditions are indicated for the Hansonburg district.

Models for Mississippi Valley-type mineralization involving two solutions which mix at the site of mineralization have been proposed (Beales and Jackson, 1976). Our fluid inclusion data does not indicate mixing of two fluids with radically differing temperature, chemistry, or salinity. Further, our geochemical analyses and calculations indicate that a two-solution model is not necessary to explain the mineralization at the Hansonburg deposit. Certainly, the regional extent of nearly identical mineralization in the Hansonburg district argues for a simple method of generating deposits.

GENETIC MODEL

Source of Mineralizing Fluids

Analyses from numerous wells in the gulf coast area (Randolph and others, 1980) are similar to analyses of inclusion fluids from the Hansonburg district (Table 2). These 90-135°C, present-day formational waters are concentrated Na-Cl brines.

Randolph and others report that when gulf coast formation-water samples were brought to the surface and cooled they precipitated numerous solid phases within the stainless-steel sample containers. The solid phases were galena, sphalerite, quartz, calcite, and halite plus an unidentified fine-grained precipitate which include major amounts of Fe, Ca, Ba, Cl, and Zn, with lesser amounts of Pb,.Cu, Ni, Mn, Cr, Cd, Mg, Sr, Si, K, Br, and S. Numerous organic compounds were also reported. These basin-brines obviously carry both metals and sulfur and precipitate sulfide minerals when cooled and depressurized. The similarity of the chemistry of the Hansonburg brines to these gulf coast formational waters, along with the hydrocarbons observed in the Hansonburg fluid inclusions, strongly suggest that the Hansonburg ore fluids were basin brines. The higher temperatures of the Hansonburg mineralizing solutions relative to the Gulf Coast formation waters can be explained by the elevated heat flow within the Rio Grande rift.

The galena-fluorite-barite deposits along the Rio Grande rift (fig. 1) are mostly open-space fillings. They have similar fluid-inclusion filling temperatures and stable-isotope compositions (Beane, 1974) as the Hansonburg deposit and possess the same general paragenesis. The Hansonburg mineralization appears to be a more extensive deposit of the

Table 2. Comparison of composition of inclusion liquids in Hansonburg district to those of other Mississippi Valley-type deposits and to present-day formation waters. [1]Hall and Friedman (1963). [2A]Roeder, Ingram, and Hall (1963)—extracted water. [2B]Roeder, Ingram, and Hall (1963)—visual estimate of water. [3]Randolph and others (1980). [4]Calculated maximum from anhydrite equilibrium. *High concentration due to mineral solubility.

	Ca (ppm)	Mg (ppm)	K (ppm)	Na (ppm)	Cl (ppm)	F (ppm)	SO₄ (ppm)
GALENA							
Cave In Rock[1]	20,600	4,000	3,100	55,400	115,000	NR	*
Wisconsin-Ill.[1]	18,500	3,400	3,200	39,000	83,000	NR	ND
Hansonburg	8,400	80	5,800	51,000	96,000	110	≤ 8[4]
FLUORITE							
Cave In Rock[1]	*	4,500	3,000	44,200	90,900	NR	3,200
Cave In Rock[2A]	*	5,600	4,800	64,600	127,100	*	17,400
Cave In Rock[2B]	*	3,600	2,900	39,500	78,400	*	7,900
Hansonburg	*	580	2,000	45,000	24,000	*	≤ 30[4]
QUARTZ							
Cave In Rock[1]	8,400	2,800	4,300	17,300	47,000	NR	12,900
Hansonburg	3,300	130	1,400	39,000	48,000	400	≤ 30[4]
FORMATION WATERS[3]							
Beula Simon	2,900	250	460	29,800	54,400	NR	380
Pleasant Bayou	8,700	650	550	36,300	79,100	NR	< 5
Fairfax-Foster-Sutter	5,500	640	820	43,700	91,900	NR	< 5

same type as those common along the Rio Grande rift. Its larger size can be attributed to a favorable host rock with much open space. We postulate that the similarity between the deposits along the Rio Grande rift is due to the presence of nearly homogeneous, Rio Grande rift formational waters which ascended along rift-activated faults and deposited mineralization in veins and cavity fillings in response to a drop in temperature and pressure. These formational-waters would be expelled by compaction, perhaps aided by local intrusive activity. They would migrate to avenues of escape, which would be major fractures associated with rifting.

Ground Preparation

A topographic high existed to the north of the Hansonburg district during Pennsylvanian time which, in all likelihood, was the result of differential uplift along the Capitan lineament. This structural and topographic high localized the deposition of the reef-facies limestone which hosts the Hansonburg deposits.

A disconformity in the sedimentary rocks overlying the Council Springs Member indicates both a depositional hiatus and an erosional surface. Karstification of the porous reef-facies most likely occurred during this period of subaerial exposure. The Council Springs section was then submerged and intermittently covered by the upper Pennsylvanian, Permian, and later formations.

Rio Grande Rift

The Rio Grande rift, since its inception, has created basins filled with sediments and downfaulted blocks capped by sedimentary rocks of Paleozoic through Cenozoic age. Formational waters within these basins could attain the 200°C temperature measured for the Hansonburg ore solutions at depths as shallow as 3.5 km as a result of the twice normal heat-flows characteristic of the rift (Reiter and others, 1975), or at even shallower depths if there was nearby volcanic activity. The formation waters would contain Pb dissolved from arkosic sediments derived from Precambrian granites. The principal source of sulfur would be Permian sediments and Permian sediment detritus. Early rift tectonics would create differential lateral displacement along the intersection of

the rift with the Capitan lineament, resulting in high-angle faults fanning from one structure to the other. These faults would intersect the formation waters at depth, providing access ways toward the surface and into the overlying sediments. The occurrence of the higher homogenization temperatures in samples taken from locations adjacent to the southeast-trending faults (figs. 2 and 5) indicates that these faults acted as conduits, allowing the ore-solutions to gain access into the cavernous Council Springs Limestone.

Mineralization

The heated formational waters migrating out of the center of the rift basin, ascending up and along fractures, would readily move laterally into the more porous strata of the karstified Council Springs Limestone. Entrance of these solutions into slightly cooler rock would result in initial deposition of quartz on the cavern walls, with later minerals (galena, fluorite, and barite) deposited as the solution cooled. As the solutions entered the cavernous host, it is likely that some of the CO would effervesce, which would further saturate the fluids with respect to galena, fluorite, and barite. The increasing relative oxidation state of the fluids with time could be due to loss of H,S by precipitation of sulfides, or simply that later, cooler solutions were less reduced.

CONCLUSIONS

The Hansonburg mineralization occurs in a Pennsylvanian reef-facies limestone. Pre-mineralization karstification along bedding irregularities created open space which was later occupied by the ore minerals. A series of southeast-trending, high-angle, reverse faults, probably related to strike-slip faulting, served as conduits for the mineralizing fluids.

Fluid inclusion homogenization temperatures range from 130-210°C and indicate the ore solutions cooled through the paragenesis. Salinities range from 10-18 eq. wt.% NaCl and were variable during the course of mineralization. Extraction and analysis of inclusion fluids indicates a solution chemistry similar to those reported for other Mississippi Valley-type deposits, and to present-day formational waters. Thermodynamic calculations based upon the analyses suggest that mineralization resulted from the simple cooling of saturated solutions.

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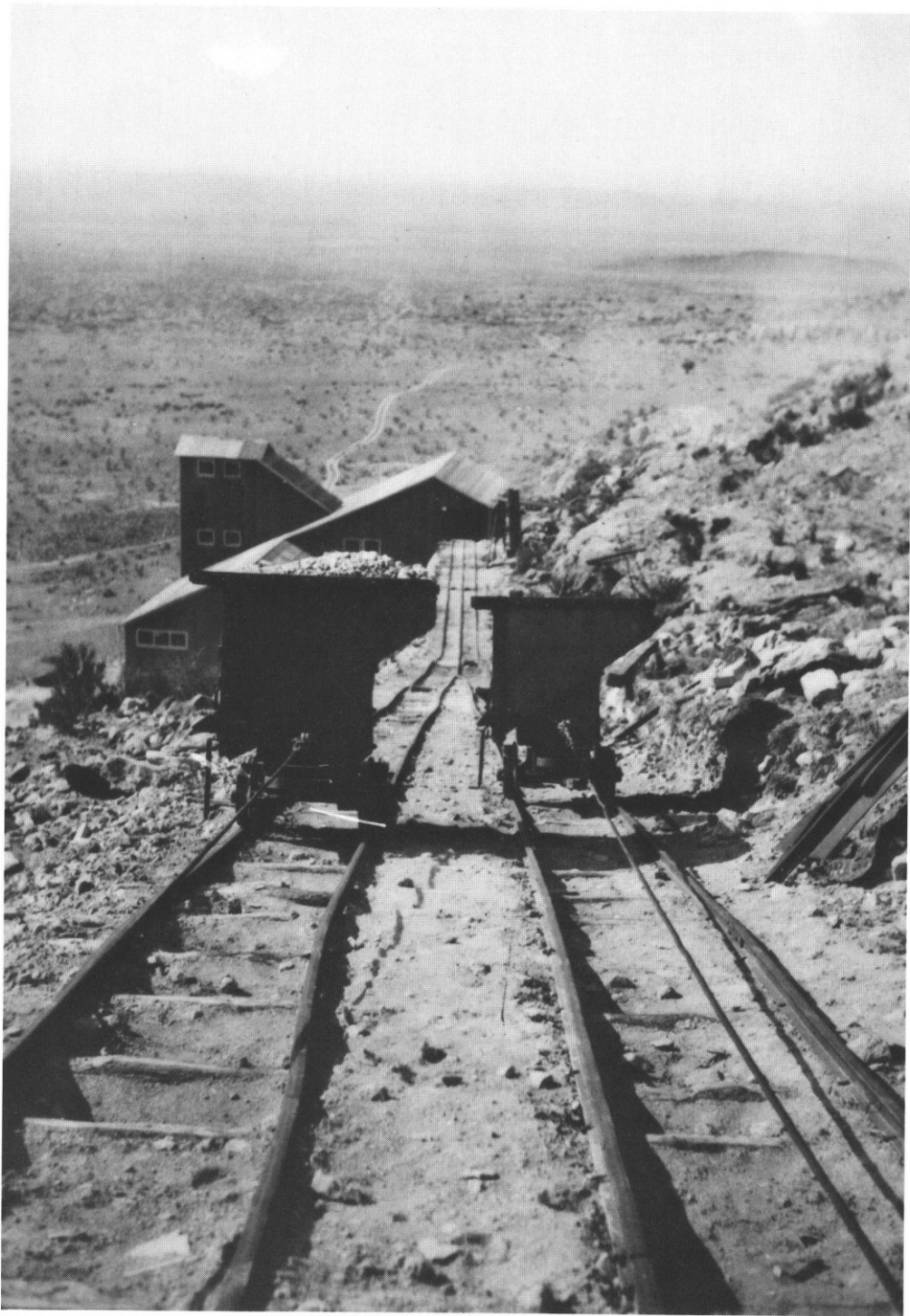
Rail tramway and mill of Western Mineral Products company, Hansonburg District, Bingham, New Mexico, ca. 1916. The Hansonburg lead deposits were discovered by the American prospectors W T. Harris and Harvey Hanson ca. 1881 but were doubtless known to the Spanish and perhaps the Indians before them. According to reports, "rough stone hammers (were) often found in the caves in the bedding plane deposits" by the early prospectors. The first serious attempt at mining was by the above company which constructed these facilities and produced several carloads of lead-silver concentrate during 1917. This and subsequent operations were unsuccessful apparently due to the scarcity of water for milling. Photo courtesy of St. Joe American Corporation, Tucson; New Mexico Bureau of Mines and Mineral Resources collection.