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Thermal Constraints on the Formation of Mississippi Valley-Type Lead-Zinc Deposits and Their Implications for Episodic Basin Dewatering and Deposit Genesis

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Abstract

The hypothesis that the Mississippi Valley-type lead-zinc deposits largely associated with basin margins formed as the result of compactive expulsion of basin brines is examined in light of two simple constraints: that the deposits formed within 1 km of the surface, and that the temperature of the ore fluids at the time of mineral deposition was 100° to 150°C. It is shown that for these near-surface temperatures to be attained the rate of fluid migration up the margins of the basin must have been more than 1,000 times greater than could be produced by the steady subsidence, sedimentation, and compaction of most basins. For base metal deposits to form as the result of the compactive expulsion of basin brines, brine expulsion must be episodic, occurring cumulatively over a period of time representing only a few thousandths of the pertinent history of basin growth. Rough estimates are made of the number of dewatering pulses a basin might undergo. If dewatering occurs in a cycle as fluid pressure builds to lithostatic pressures and is then released, 50 dewatering pulses occur, approximately one every million years, as a strata subsides from 3 to 5 km depth. At 3- to 5-km depths temperature is in the range of Mississippi Valley-type mineralization and thus strata at these depths could have provided the fluids that deposited ore. Episodic pulses of geopressured brines moving through the near-surface sites of ore deposition, followed by much longer periods of exposure to cool surface waters, could produce the color banding of sphalerite observed in the Upper Mississippi Valley district, the cycles of sulfide precipitation and dissolution observed in most Mississippi Valley-type deposits, and the distinctive local tectonic features associated with Mississippi Valley-type deposits. Shale membrane filtration would occur while the fluids are held in a geopressured state prior to rupture and sudden venting. This process would increase the salinity of the pore fluids. Basin features favoring the right kind of episodic dewatering are identified. These criteria appear to be useful for identifying basins with associated lead-zinc mineralization.

Introduction

MISSISSIPPI Valley-type massive sulfide deposits are stratiform deposits of lead or zinc sulfide and gangue (calcite, quartz, dolomite, flourite, barite) that precipitated epigenetically in open space (often around breccia fragments in caves or collapse structures) from moderate temperature brines (15–30 equiv. wt % NaCl). Fluid inclusions indicate the temperatures during ore deposition ranged from 80° to 200°C but were most commonly 100° to 150°C (Anderson and Macqueen, 1982). The deposits are associated with pinch-outs that surround domal structures, basin growth faults, and near-shore reefs. Most deposits occur in limestone or dolostone, but some have sandstone hosts (Ohle, 1952; Beales and Onasick, 1970). Most are located at the edges of sedimentary basins; some are associated with faults located over the central parts of basins (see Figs. 1 and 2). Although economic metal concentrations are unusual, occurrences of lead-zinc mineralization appear to be almost ubiquitous in and around some basins. The lead-zinc mineralization is not, in general, related to igneous activity, although this may be the case in a few instances. The primary association is to basins. Some basins appear to have much more associated lead-zinc mineralization than others (Anderson and Macqueen, 1982).

These general observations have lead most economic geologists to conclude that Mississippi Valley-type deposits formed as the result of a normal basin process—namely, the expulsion of briney pore fluids from basin strata as the basin subsides, thickens, and compacts. In this view, pore solutions squeezed out of the basin migrate at least in part downward to the warmer portions of the basin and then laterally along

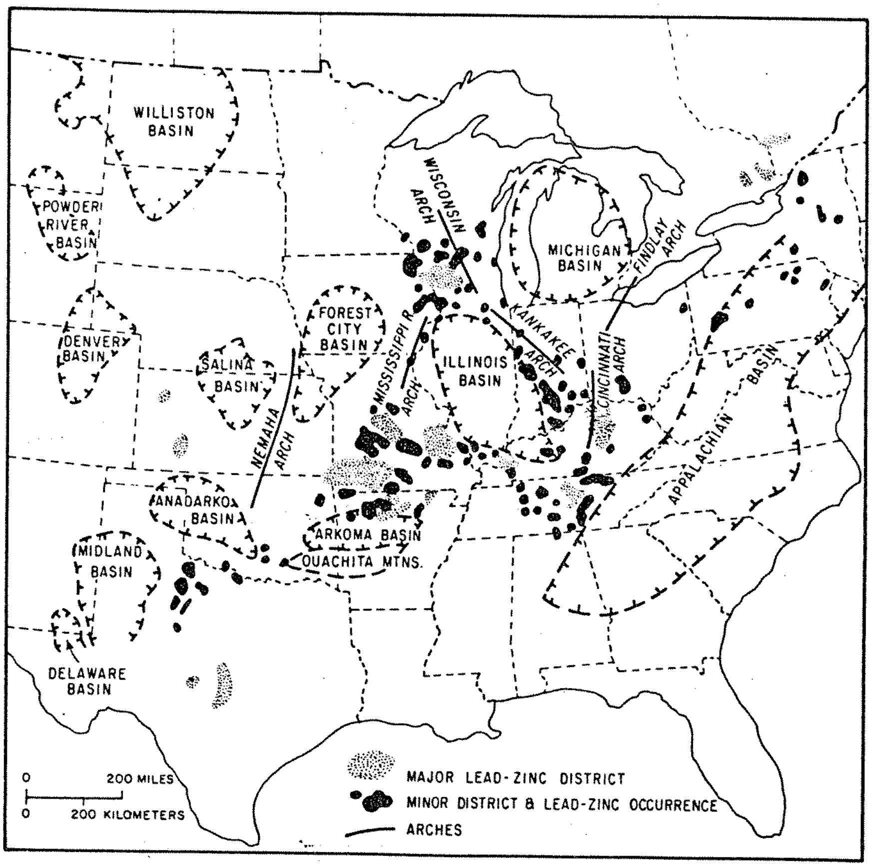


FIG. 1. Mississippi Valley-type lead-zinc deposits and occurrences are ubiquitiously distributed around some basins. Simplified from figure 1 of Anderson and Macqueen (1982).

a basal aquifer to locations where the aquifer crops out or where growth faults or reefal structures provide cross-strata channels from the aquifer to the surface.

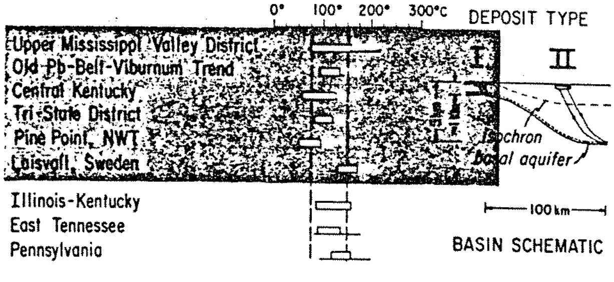


FIG. 2. Fluid inclusions in sphalerite from various Mississippi Valley-type deposits have filling temperatures in the range 75° to 150°C. In this paper we classify the deposits into two subtypes: one occurring at the edges of basins (as indicated by the shaded block), and one associated with faults in the centers of basins. This paper considers only deposits of the first, more common, type. Fluid inclusion data shown in this figure were taken from the following references: Howe, 1981; Rickard et al., 1979; Roedder, 1971, 1976, 1977.

In this view, ore deposits form when escaping solutions channel through a particular escape locality. Lead-zinc occurrences of subeconomic grade form where the discharge is less focused. The warm temperatures are obtained from the deeper parts of the basin and are carried upward by the expelled fluids. Near the surface the metals carried in the brine precipitate due to cooling or mixing with other solutions carrying sulfide sulfur.

How near the surface precipitation occurs is uncertain. Stratigraphic evidence suggests it occurs within a few hundred meters to a kilometer of the surface (Anderson and Macqueen, 1982; P. Gerdemann, pers. commun.). The basinal brine-pore fluid expulsion model was first proposed by Noble (1963) and has since been argued and extended most notably by Beales and Jackson (1966), Jackson and Beales (1967), Beales and Onasick (1970), and Dozy (1970). The reader is referred to these papers and an excellent review by Anderson and Macqueen (1982) for further elaboration and documentation of the summary statements made above. There is much that we do not understand. Reviews by Ohle (1980) and Anderson and Macqueen (1982) highlight this.

Dozy (1970) pointed out that the expulsion of pore fluids, in addition to occurring steadily with gradual basin compaction, could occur in bursts associated with the rupture of pockets of abnormally pressured (geopressured) fluid in low permeability units in the basin. Such a "settling flood," escaping from a zone of near-lithostatic fluid pressure, could "lift" or "float" the covering section. Settling floods could cause many of the small deformation features observed in almost all Mississippi Valley districts such as "small scale folding or crumpling and small faults or shears limited to certain strata or between adjacent layers of different competence," pitch and flat features, gash veins, and sandstone dikes (Dozy, 1970, p. 167).

Recently it has become clear that ore was not deposited all at once but rather was deposited in many pulses between which the precipitated ore minerals were often partly redissolved. Perhaps the most spectacular manifestation of this pulsed nature of mineralization is the color banding in sphalerite in the Upper Mississippi Valley district elegantly documented by McLimans et al. (1980). The ore deposits of the Upper Mississippi district consist of a 4-cm coating of sphalerite on breccia fragments. The sphalerite is color banded due to variation in iron content and is easily seen in polished thin sections. There are about 30 distinct dark marker bands in the district, and as McLimins et al.'s (1980) illustrations eloquently show, the band pattern can be easily correlated over distances as great as 23 km. Minor bands within the major bands may or may not correlate on a district scale. These bands could be due to local hydrologic variations. When examined in detail it is observed that many of the bands were corroded before the next band was deposited. Thus, sphalerite deposition proceeded in pulses of deposition and dissolution. Similar cycles of deposition and dissolution are observed in most districts (Ohle, 1980). Sverjensky (1981) musters compelling paragenetic and textural evidence to show that such cycles occurred in the Buick mine of the Viburnum Trend, for example. The dissolution of multiple stages of galena is clearly evident in SEM photographs of samples from the Buick mine (Sverjensky, 1981) and can be clearly seen in hand specimens on display in the mine offices of other mines in the Vibernum Trend (for example, Saint Joseph Mining Company, Vibernum No. 28 mine). The cycles of dissolution and precipitation have been generally attributed by the above authors to relatively minor fluctuations of solution chemistry. In this paper we will argue they must be due to much more major pulses of basinal fluid expulsion and much more dramatic cycles in fluid temperature than previously imagined.

There has been only one attempt that we are aware of to reduce the conceptual basinal brine-pore fluid expulsion model to a quantitative mathematical form. Sharp (1978) showed that geopressured zones would develop over 40 m.y. of sedimentation in a basin with sediments of 0.05-mD permeability if the sedimentation rate at the center of the basin was 250 m/m.y. and 25 m/m.y. at the edge. Sharp further showed that, if the permeability of the basin were increased to 300 mD by faulting, enough fluid would migrate downward to a 30-m-thick, 30 D basal aquifer, and along the aquifer to where it cropped out 200 km from the basin edge, to form easily a Mississippi Valley-type deposit at the location of the aquifer outcrop. In Sharp's preferred model the fluid drained in ~1,000 years and the Darcy velocity of the fluid at the area of outcrop of the aquifer was ~1 m/yr.

Sharp did not consider the conditions under which flow in the basal aquifer up the side of the basin could carry heat from the deep central portions of the basin to near the surface. Most have probably assumed, as we did initially, that there would be little difficulty in this regard. However, such intuitive perceptions have proven to be entirely false. The requirement that fluids squeezed out of a basin by compaction be able to carry the temperatures of the deeper portions of the basin to within 1 km of the surface (where the solutions could enter a growth fault or a high permeability reefal structure) turns out to be a surprisingly restrictive condition. It means, in fact, that if the basinal brine-pore fluid expulsion model is to work, the expulsion of brines from a basin must be an episodic phenomenon, occurring cumulatively over only a small fraction of the total time of basin development.

The first part of this paper is devoted to a demonstration of the above statement. We show that for reasonable rates of basin subsidence and compaction the fluid flow out a basal aquifer will not perturb the normal geothermal gradient at all. Perturbation of the thermal gradient requires flow rates at least 300 to 5,000 times those obtainable by steady subsidence and compaction, even assuming all the fluid squeezed out of the basin finds its way down to and flows along the basal aquifer to the surface.

In the second part of the paper the number and nature of pulses of fluid expulsion that might occur in a basin are estimated and basin characteristics favorable for Mississippi Valley-type mineralization are identified. Many of the seemingly odd characteristics of Mississippi Valley-type deposits, including the cycles of ore deposition and dissolution and sphalerite banding, and the apparent coincidence of mineralization, brecciation, and minor deformation (Ohle, 1980), can be naturally accounted for by pulses of fluid expulsion, and many are very difficult to understand or account for in other ways.

Basin characteristics conducive (from a theoretical fluid flow-heat balance point of view) to Mississippi Valley-type mineralization are in some cases new and might be useful exploration guides. The final section of this paper compares the characteristics of basins that have produced Mississippi Valley-type mineralization, and basins that have not, to the characteristics that seem to be important from the basinal brine-pulsed fluid expulsion viewpoint.

Can the Flow Caused by Normal Basin Compaction Move 100° to 150°C Brines to within 1 km of the Surface?

The first questions are: what magnitude of fluid flow along a basement contact would be required to elevate the basement temperatures within 1 km of the surface from normal values of about 50°C to 100°–150°C, and could this magnitude of flow be provided by steady basin compaction? To answer these questions we need to know the typical slope of the sides of basins, the typical size of the basins associated with Mississippi Valley-type deposits, their rate of subsidence and compaction, and the basal heat flow into the basins as a function of their age.

Figure 3 shows several cross sections through deposits in the mid-continent Mississippi Valley area. The basins associated with the deposits are typically ~600 km wide and the slopes of the basin margins are generally less than 1 percent (Fig. 2). The deposits, except for the Illinois-Kentucky district deposits and some other deep fault-controlled districts (all hosted by Mississippian-age rocks), are located at the edges of the basin.

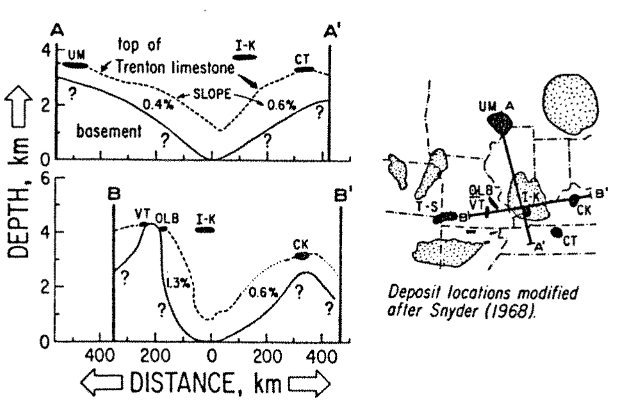


FIG. 3. Cross sections through Mississippi Valley-type deposits in the midcontinent region indicate that basins which may have formed the deposits were at most 600 km wide and had marginal slopes generally less than 1 percent. The cross sections were prepared from the Tectonic Map of the United States, 1962.

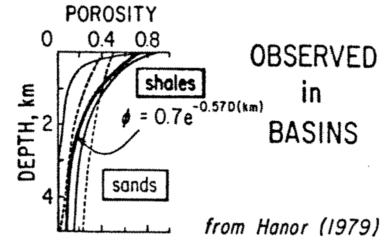


FIG. 4. Basin strata are observed to have porosities that decrease with the depth of burial of the strata. General porosity versus depth relations are reproduced from Hanor (1979). The porosity versus depth relation used in this paper is shown by the dark line in the shale field, and the equation that describes this line is given.

The amount of fluid expelled from a basin as it grows can be deduced from the rate of basin subsidence and an estimate of the porosity as a function of depth in the basin. The latter has been well documented by work related to oil exploration and recovery and is summarized in Figure 4 (Hanor, 1979). Sediments and rock strata in basins are observed to have porosities that depend mainly on depth. The conclusion is that, as these strata are loaded by overlying material, they are compacted and their porosity reduced. This happens whether or not the sediments are indurated. As shown in the figure, sands compact somewhat differently from shales and there is a fairly broad range of compaction versus depth spanned by sediments and rocks from different basins. The curve we use in the calculations that follow is shown as the solid line within the shale field in Figure 4. The figure gives the exponential function that describes this line, i.e., the function which gives porosity as a function of depth that we shall use below.

The mass of fluid that is squeezed out of a unit volume of sediment, Q, can be simply described:

$$Q\ [\mathrm{g/cm^3 - sec}] = \frac{\partial(\rho\phi)}{\partial t} = \frac{\rho \partial z}{\partial t}\frac{\partial \phi}{\partial z} = \rho v_z \frac{\partial \phi}{\partial z}. \quad (1)$$

We have assumed that pore fluid is incompressible (i.e., that the fluid density is constant) and used the chain rule to convert the change in porosity, ϕ, with time to the product of the rate of subsidence of the basin, v_z, and the change of porosity with depth. The change of porosity with depth is, of course, just the function determined in Figure 4. In this simple model we do not consider the decrease in thickness of rock strata as compaction takes place. We simply seek a reasonable estimate of the rate at which pore fluids can be squeezed out of basin sediments if this process takes place steadily as the basin subsides and grows.

We can integrate Q over the depth of the basin and ask how much fluid a 600-km-wide basin might expel through basal aquifers at its sides. Considering the basin to be two dimensional (i.e., very long compared to its width), and assuming fluids will be expelled equally to each side, the maximum total rate of fluid expulsion up a marginal basal aquifer per cm length of the two-dimensional basin, M_{300}, will be:

$$M_{300}\,[\text{g/cm-sec}] = 300 \times 10^5 \int_{z=0}^{z=D} Q\,dz. \quad (2)$$

M_{300} is the total amount of fluid expelled from a 1-cm cross-sectional strip across a basin 300 km wide (half the width of the whole basin) and D km deep.

To complete the model we need estimates of both the depth of the basin, D, and the basal heat flow into the basin, A, as a function of time. We obtain these parameters by analogy to the conductive cooling and subsidence of the oceanic plate at mid-ocean ridges, corrected for sediment loading. Kinsman (1975), Sleep (1976), and Sclater and Christie (1980) among many others have argued that this is a reasonable model for the subsidence of continental margins, and it is a reasonable starting place for modeling of epicontinental basins as well. Thus we take:

$$D = 7.36(1 - e^{-t/62.8}) \quad (3)$$

and

$$A = 11.5/\sqrt{t}, \quad (4)$$

where A is the basal heat flow in Heat Flow Units (HFU; cal/cm^2-sec $\times 10^{-6}$) and t in both formulas is time in millions of years.

Equations (1) to (4), plus the porosity versus depth relation shown in Figure 4, represent a simple model of basin compaction. The precise appropriateness of elements of this model could certainly be debated, but no single element is probably incorrect by more than a factor of two for most basins. The conclusions we will draw from the model are insensitive to this magnitude of model uncertainty. The important point is that Mississippi Valley deposits were thermal anomalies at the time of deposition and that we now have a handle on the rate at which a basin can expel fluids under quasi-steady state conditions and are in a position to inquire whether this rate of fluid expulsion could produce the anomalous near-surface temperatures indicated by Mississippi Valley-type mineralization.

Table 1 evaluates the model for basins of three different ages. For each age basin Table 1 gives the depth of the basin, the temperature and porosity of the deepest parts of the basin (assuming 20°C near-surface temperature), the rate of subsidence of the basin (obtained by differentiating equation (3) with respect to time), and the rate of fluid expulsion from a 1-cm cross-sectional strip across the 300-km half-width of the basin (M_{300}). The table indicates that a basin must be at least ten million years old to attain basal temperatures in excess of 100°C. Although the heat flow is greater for younger basins, younger basins are not deep enough to have basal temperatures in the range indicated by most fluid inclusions associated with Mississippi Valley-type mineralization. The rate of fluid expulsion increases to a maximum at about 40 m.y. and then decreases. The reasons for this dependence are similar. At first the basin is subsiding rapidly, but there is not a great thickness of sediments accumulated from which to squeeze fluids. Later there is a great thickness of sediments, but the rate of subsidence and compaction has slowed. In between is the maximum rate of fluid expulsion. The situation is different in some geological cases such as the Michigan basin which had two episodes of subsidence and thus had an unusually thick sediment pile during the early (rapid) part of the second subsidence. In this case the fluid expulsion rate might have been three times the rate given in Table 1 for the 10-m.y.-old 1-km-deep basin (because from the point of view of contained pore fluids and compaction the basin was 3 rather than 1 km deep). Our point is not that Table 1 is exactly accurate in all cases but rather that it gives a reasonable base estimate for the rate of dewatering of basins that is unlikely to be in error by more than a factor of three or so.

Assuming that all the fluid squeezed out of the basin finds its way to the basal aquifer, would the flow up the sides of the basin be sufficient to perturb temperatures near the surface significantly? Figure 5 shows the problem schematically; Figure 6 shows the method used to calculate the perturbation of temperature at the sides of the basin. We use a simple forward finite difference scheme and assume thermal steady state. The question we ask is whether it is possible, after fluid has been flowing in the basal aquifer for a long time, for warm fluid from the deep parts of the basin to reach the near-surface without cooling substantially. We assume that the fluid flows in a fairly thin basal aquifer. A thicker aquifer would diminish the thermal perturbation of the basin.

TABLE 1. Expulsion Rates for Basins of Different Ages

Age (m.y.)	D (km)	T(D)[1]	ϕ(D)	v_z $\left[\frac{m}{m.y.}\right]$	M_{300} km [g/cm-sec × 10^3]
10	1.1	93°	0.37	100	3.1
40	3.5	147°	0.09	62	3.6
80	5.3	156–179[2]	0.03	33	2.1

[1] K = 5 × 10^{-3} cal/cm-sec-°C
[2] 1.3 to 1.5 HFU

Equations (1) through (4) and the porosity versus depth relationship of Figure 4 are used to calculate the depth, D, basal temperature, t(D), basal porosity, ϕD, rate of subsidence, v_z, and rate of fluid expulsion from a 300-km-long, 1-cm-wide cross-sectional strip across the basin M_{300}; K is the thermal conductivity of the basin assumed in the calculations

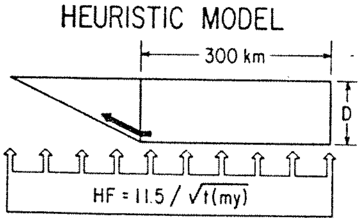
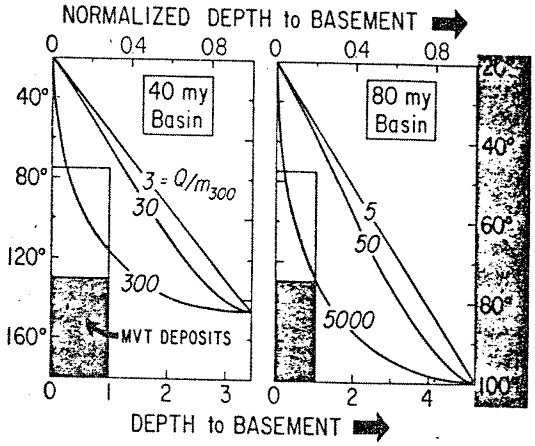


FIG. 5 A diagrammatic portrayal of the situation calculated in Figure 6. Basal heat flow (in heat flow units) depends on the age of the basin. All fluids compactively expelled from a 300-km-wide (half-width) basin are assumed to be expelled downward and up the margin of the basin along a basal aquifer.

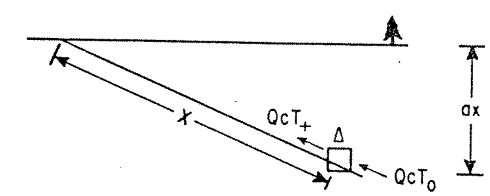


	left scale	right scale	
K	5×10^{-3}	10×10^{-3}	cal/cm-sec-°C
	2.1	4.2	watts/m-°C
TEMPERATURE at TOP of BASEMENT			

FIG. 7. Thermal profiles along the margins of a 40- and 80-m.y.-old basin are calculated assuming a 1 percent margin slope using the methods outlined in Figure 7. The heat flow in the base of the 40-m.y.-old basin is 1.82 HFU and for the 80-m.y.-old basin 1.5 HFU. The rate of fluid outflow is described in the diagrams in terms of the ratio of the rate of outflow per unit length of the basin margin used in the calculations, Q, to the rate of fluid outflow per unit margin length that could be produced by steady basin compaction, M_{300} (see values in Table 1). Results are shown for two different values of basin thermal conductivity, K. The higher value of thermal conductivity (right-hand scale) is probably more appropriate for shale-rich basins, the lower value more appropriate for shale-poor basins. The diagrams show that fluid expulsion rates 300 to 5,000 times those that could be provided by steady basin compaction, with all the expelled fluid exiting up the basal aquifer, are necessary to raise the temperature within 1 km of the surface to 75° to 150°C. This means that if compactively expelled brines are to be the ore fluids of Mississippi Valley-type deposits, the brine expulsion must occur infrequently and episodically, at many thousands of times the average rate.

The results are shown in Figure 7 for basins 40 and 80 m.y. old. The diagrams in Figure 7 show temperature as a function of depth to the basement contact at the side of the basin. The temperature axis is given for two different values of the average thermal conductivity of the basin. The left-hand axis is probably more appropriate for shale-rich basins, and the right-hand axis more appropriate for shale-poor basins (Touloukian and Ho, 1981; Clark, 1966, p. 462). A still lower thermal conductivity, appropriate for porous sediments (Sclater and Christie, 1980), would show similar curves but with hotter temperatures on the ordinate scale. The figures show that, for fluid expulsion rates even three to five times greater than could be steadily maintained by the basin, the temperature profiles remain unperturbed from their normal linear increase with depth. Aquifer flow rates 300 to 5,000 times those sustainable by quasi-steady state compaction are required for the temperature profiles to intersect the shaded boxes at the left of the diagram and account for the perturbed temperatures of Mississippi Valley deposits. These boxes represent the conditions required by fluid inclusions associated with Mississippi Valley-type mineralization—that 75° to 150°C fluids reach within 1 km of the surface. For such temperatures to be attained near the surface, especially the 100° to 150°C temperatures commonly recorded in fluid inclusions in the deposits, ratios of

TEMPERATURE DISTRIBUTION FOR FLOW UP A BASEMENT CONTACT



$$-Qc\frac{\partial T}{\partial x} - \frac{K}{ax}T + A = 0$$

$$-Qc\frac{(T_+ - T_0)}{\Delta} - \frac{K}{ax}\frac{T_+ + T_0}{2} + A = 0$$

$$\boxed{T_+ = \frac{T_0\left(\frac{Qc}{\Delta} - \frac{K}{2ax}\right) + A}{\frac{Qc}{\Delta} + \frac{K}{2ax}}}$$

FIG. 6. A simple forward finite difference scheme is used to calculate the steady state temperature distribution along the margin of the basin. The total flow per cm length of the margin, Q, is assumed concentrated in a fairly thin aquifer. A is the heat flux in cal/cm²-sec-°C, a is the slope of the aquifer, K is the thermal conductivity of the basin in cal/cm-sec-°C, c is the heat capacity of the pore fluid in cal/g-°C, and Q is measured in g/cm-sec. The results of calculations of the type shown are given in Figure 7.

Q/M_{300} of 300 to 5,000 or more are required. It is remarkably difficult to perturb the near-surface temperatures. Such perturbation depends on the rate of fluid expulsion from the basin and not the thermal conductivity of the basin sediments; Mississippi Valley deposits are not associated with basin conductivity anomalies.

We believe Figure 7 indicates that the existence of Mississippi Valley-type deposits and occurrences requires abnormally rapid expulsion of basin brines. The model is approximate, but we have been generous in assuming all fluid is expelled downward to the basal aquifer. We have assumed thermal steady state. This is conservative since transient thermal losses will be greater than the steady state losses (the transient heating up of rock along the fluid path represents an additional heat sink). Transient heat losses will make it more difficult for warm basal fluids to reach the near-surface. Channeling or focusing in the basal aquifer is to be expected, and could increase fluid velocities. The wide distribution of lead-zinc mineralization around a basin (see Fig. 1) does not suggest channeling of the magnitude required (×300 to 5,000).

Exothermic reactions involving the dewatering of clays could contribute heat. The enthalpies of reaction are quite uncertain, but even taking the upper range of 800 kcal/mole, the heat contributed by dewatering of 10 wt percent basin clay would be equivalent to only 0.1 HFU—hardly a significant increase in basal heat flow (see Fig. 8). Exothermic dewatering reactions could contribute very significant heat if they occurred episodically, however, and this possibility should be kept in mind.

The above considerations indicate quite clearly that the near-surface temperatures recorded in fluid inclusions in Mississippi Valley deposits could not have been produced by the outflow of basin fluids resulting from steady basin compaction. Two alternatives present themselves: either the basinal brine-pore fluid expulsion hypothesis must be abandoned, in favor of some quite different hypothesis, or pore fluid expulsion must be an episodic phenomenon that occurs only over a very small fraction ($\lesssim$1/1,000th) of the relevant history of basin accumulation.

EXOTHERMIC CLAY DEWATERING

- Montmorillonite → Illite (3-4 Km Depth)
- ΔH_R = -50 to -800 kcal/mole
  $\cong$ -125 to -2000 cal/g
- [montmorillonite] $\gtrsim$ 10 Wt %

$$HF_R = v_z \,[\mathrm{Mont}]\, \rho_r \Delta H_R$$

T [my]	HF_R
40	.006 to .1 H F U
80	.003 to .06 H F U

FIG. 8. Clay dewatering reactions could contribute heat to warm expelled fluids. If clay dewatering occurs steadily as the basin grows and subsides, the amount of heat that could be provided by this mechanism is minor. At most the heat flow from the basin would be raised 0.1 HFU and this would lead to an insignificant increase in near-surface temperature (see Fig. 10). (Enthalpies of reaction are estimated from data in Helgeson, 1969; Robie et al., 1978; Nriagii, 1975; Boles, 1981; Nesbitt, 1977.)

The Possible Nature of Episodic Pore Fluid Expulsion and Basin Characteristics Favorable to Mississippi Valley-Type Mineralization

Largely because Mississippi Valley-type deposits are so clearly associated with basins, and so poorly associated with anything else (i.e., igneous intrusions), the second alternative, above, deserves close attention. In this section we attempt to estimate the nature and frequency of episodic basin dewatering and use these estimates to identify those basin characteristics favorable to Mississippi Valley-type mineralization.

The basic approach used is illustrated in Figure 9, which shows finite difference calculations of the thermal effects of compaction-driven fluid flow in a 40-m.y.-old, 3.5-km-deep, 400-km-wide basin (200-km half width). The first cross section shows the relationship between basement and basin strata. The permeability is maximum at the base of the basin and along the basin-basement contact. It drops off strongly with depth into the basement and upward into the basin.

The first cross section at the upper left shows the temperature distribution in the basin after 1 m.y. of steady fluid outflow. The fluid source term is calculated from the rate of basin subsidence appropriate for a 40-m.y.-old basin and the porosity versus depth as shown in Figure 5 and as discussed in the Appendix. Fluid flow does not perturb the temperature profiles at all. The exothermic dewatering of clay elevates the temperature contours slightly on the basin side. Eight hundred kcal/mol, 10 wt percent clay, and complete clay dewatering between a 2- and 3-km depth in the basin is assumed (see Fig. 8). Fluid flow is shown by the contours of constant vertical fluid velocity. The contours along the basement contact thus indicate flow.

The other cross sections show the effects of fluid expulsion when one million year's worth of fluid is expelled in a shorter period of time—100,000, . . . 10,000, . . . down to 10 years, for example. Such expulsion might occur as the result of episodic compaction. The sections show that episodic venting can

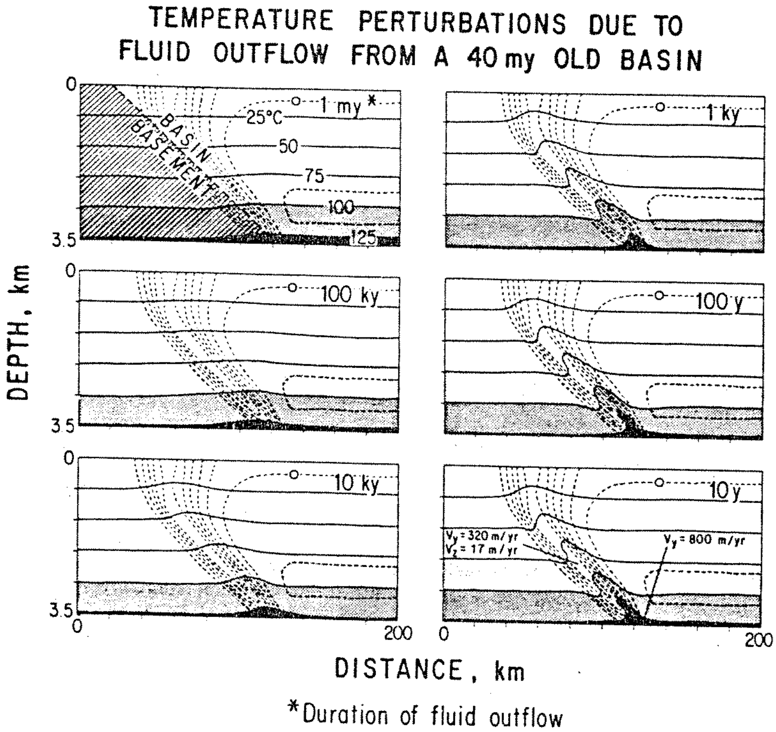


FIG. 9. The effect of increasing rates of fluid expulsion on thermal profiles in a basin. Finite difference techniques are used to solve the coupled fluid flow and heat balance equations as discussed in the appendix. Permeability in this 200-km half-width basin decreases strongly both upward and downward from the basin-basement contact. (It decreases one order of magnitude for each 10 grid points up or down. The computational grid is 40 × 30.) The basin-basement contact angles upward at a slope of ~4 percent in the left-hand half of the diagram, as shown in the first section. Contours of vertical (Darcy) fluid flow show the flow is most intense along the lower parts of the basal aquifer. The rates of fluid expulsion for the first section in the figure are appropriate for the porosity decrease with depth (equation in Fig. 4) and the rate of subsidence of a 40-m.y.-old basin (Table 1). The temperature distribution is seen to be unchanged after 1 m.y., despite the fluid expulsion. The temperature is elevated slightly on the basin side due to the dewatering of 10 wt percent clay between 2 and 3 km depth (800 kcal/mole heat of reaction assumed). Subsequent figures show the effects on thermal profiles of dewatering 1 million year's worth of compactive, expelled fluids in shorter periods of time. The rates of fluid flow in these figures are correspondingly greater, although the pattern is unchanged. The contour values are shown for the last figure in m/year Darcy velocity. The rates of flow are most rapid in the last figure and are one order of magnitude smaller in each preceeding figure. The figures show that fluids must be expelled at 100 to 1,000 times the normal rate before temperature distributions in the basin are significantly perturbed, but that after a point, further decreases in the time of dewatering do not further increase the thermal perturbation. This observation allows definition of the volume of dewatering pulse required to expel warm fluids to the surface, as detailed in the text and shown in Figure 11.

perturb the temperature along the basement contact but that, after a point, further increases in the rate of venting do not lead to greater temperature perturbations. The temperature perturbations for the 100- and 10-year venting episodes are identical, for example.

The reason arbitrarily large thermal perturbations cannot be produced is, of course, that a limited quantity of fluid is expelled. Once flow is rapid enough to minimize conductive losses, this limited mass of fluid can carry the thermal anomally only so far. Had a greater fluid mass been vented (rather than just 1-m.y.'s worth of compaction), the thermal anomaly would have extended farther and might even have reached the surface. We thus see there is a relationship between the mass of fluid in an expulsive pulse, the characteristics of the basal aquifer, and the degree of temperature perturbation along the aquifer. If we could estimate the fluid mass expelled in a single pulse, we could state the maximum thickness of basal aquifer which would allow significant thermal perturbations to reach within 1 km of the surface.

It may be worth while at this point to summarize our approach in terms of a simple analogy. Consider a hot water tank connected to a spigot by a long pipe. In the first part of this paper we considered whether hot water could be drawn from the spigot if the capacity of the tank were infinite and the spigot left open for long enough for the system to come to thermal equilibrium. We showed in Figure 7 that the flow rate had to be much larger than that which could be supplied by steady basin subsidence (and quasi-steady state compaction) for the conductive losses along the pipe to be balanced by the flow of heat along the pipe and for hot water to be delivered out the spigot. Now we consider the finite capacity of the tank. How much of the pipe can be heated by discharge of the tank? Clearly if the capacity of the tank is small it may not be possible to heat up the pipe sufficiently for any hot water to be delivered out the spigot. This problem is illustrated in Figure 9 and discussed immediately above. The pure transient part of the problem (which we consider in a moment) assumes the pipe is perfectly insulated (but initially cold) and asks how much hot fluid is required to warm the pipe so hot water can be delivered out the spigot.

One approach to estimating the mass of fluid in a single expulsive pulse is to assume that a geopressured zone will dewater when the fluid pressure reaches the lithostatic or overburden pressure. The fluid pressure in geopressured zones generally ranges up to but does not exceed lithostatic values (Jones, 1975). When lithostatic pressure is obtained, the geopressured zone is assumed to rupture and drain, dropping fluid pressures back down to normal hydrostatic values. At this point, continued basin subsidence and sediment load accumulation recompacts the formerly geopressured zone, bringing fluid pressures back to lithostatic, whereupon another expulsive rupture occurs.

The rock matrix is much more compressible than the pore fluid, and if the host is totally impermeable, the increase in fluid pressure due to a porosity decrease may be expressed:

$$\Delta p = \frac{1}{\alpha}\frac{\Delta V}{V} = \frac{1}{\alpha}\frac{\Delta\phi}{\phi} = p_{\mathrm{lithostatic}} - p_{\mathrm{hydrostatic}} = (\bar{\rho}_R - \rho)gD. \quad (5)$$

Here $\bar{\rho}_R$ is the average rock density above the geopressured zone, α is the compressibility of water, V is the volume of water, p is the fluid pressure (lithostatic or hydrostatic), g is the gravitational constant, and D the depth of the geopressured zone. $\bar{\rho}_R$ may be expressed:

$$\bar{\rho}_R = \bar{\phi} + (1 - \bar{\phi})2.7, \quad (6)$$

where $\bar{\phi}$ is the average formation porosity above the geopressured zone, which may in turn be expressed:

$$\bar{\phi} = \frac{1}{D}\int_0^D \phi \, dz \quad (7)$$

where $\phi(z)$ is given by the expression in Figure 4. The density of the pore fluid is assumed to be unity in equation (6). We can now calculate the increment in ϕ required to increase fluid pressure from hydrostatic to lithostatic. We augment ϕ by this value to determine the value of ϕ at which rupture of the geopressured zone occurs. The inverse of the porosity versus depth relation in Figure 4 is then used to determine the depth of the geopressured zone at the time of rupture, and given this depth, the inverse of equation (3) is used to determine the age of the basin at the time of rupture. The $\Delta\phi$ required for the next rupture is then determined from equation (5), and the series of calculations repeated. Thus, through a cyclic series of calculations a hypothetical history of rupture of an impermeable strata in a basin can be determined. The cyclic calculations are shown below in equations (8) to (11). $\bar{\rho}_R$ is, of course, updated using (6) and (7) at each cycle.

$$\Delta\phi = \alpha\phi(\bar{\rho}_R - \rho)gD, \quad (8)$$

$$\phi = \phi + \Delta\phi, \quad (9)$$

$$D = -1.75 \ln \phi/0.7, \quad (10)$$

$$t = -62.8 \ln (1 - D/7.36). \quad (11)$$

The results of the above calculation scheme are shown in Figure 10. The figure shows the change in porosity during each expulsive episode (right-hand axis and heavy solid curve) and the time interval between each expulsive pulse (left axis and lighter curves) as a function of the depth of the geopressured zone. The number of expulsive episodes that might occur in strata starting when the strata is at 1-km depth is shown along the lighter curve. There are hundreds of episodes of expulsion, but between the depths of 3 and 5 km, where the temperature in the deep parts of the basin would be great enough to produce Pb-Zn mineralization, the number of expulsive episodes is the difference between 187 and 138 or about 50. The time between expulsive episodes for geopressured zones at a 3- to 5-km depth is about 1 m.y., and the porosity change associated with fluid expulsion is about 0.1 percent.

These numbers are very rough estimates, as will be discussed below, but do provide some basis for illustrating a further constraint on basin characteristics that is required for a dewatering pulse to carry deep basin temperatures to the surface. This is illustrated in Figure 11. If we assume the basin pore fluids are expelled up the basal aquifer fast enough so that conductive heat losses to the confining strata are minimized, we are left with the minimum requirement that the fluid pulse contain at least enough heat to warm the rock contained in the basal aquifer to the temperature of the deep part of the basin. Since we

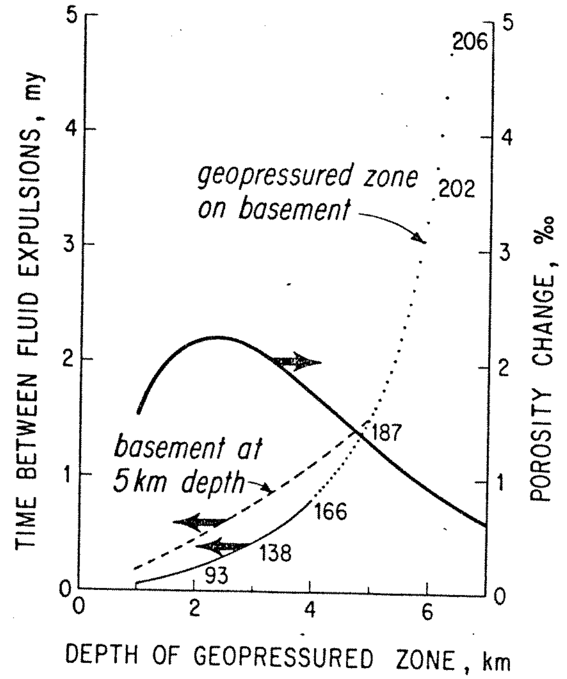


FIG. 10. Time between episodes of dewatering of geopressured zones (dotted curve) and porosity drops associated with dewatering (heavy solid curve). This is calculated assuming dewatering occurs as part of a cycle in which normal matrix compaction builds pore pressure to lithostatic values, causing rupture and drainage of hydrostatic fluid pressures; pressure then rebuilds to lithostatic values and followed by rupture, etc. Both curves are shown as a function of depth to the geopressured strata, which is assumed (except for the dashed curve) to lie at the bottom of the basin. For the dashed curve, basement is assumed to lie at a 5-km depth and the geopressured zone at higher elevations. Geopressured zones above basement dewater somewhat less frequently because the rate of subsidence is less at any given depth. The difference between the dashed- and solid-dotted detail curve is not great. Numbers along the solid-dotted time interval curve indicate the number of dewaterings that have occurred in the strata in moving from 1 km to various greater depths. Dots indicate individual episodes of dewatering. The number of dewatering episodes between the 3- and 5-km depths when the strata was warm enough to provide ore fluids, for example, is 49. The porosity change associated with dewatering in this interval is a few tenths of a percent.

want warm waters to vent and deposit lead-zinc mineralization, we write the condition that the heat provided by the deep brines exceed the heat required to warm the basin margin portion of the basal aquifer by at least a factor of two. After some algebraic manipulation we conclude, as shown in Figure 11, that this condition can be met provided that the basal aquifer is thinner than some value that is given by a collection of parameters describing the magnitude of the fluid pulse (D, W, $\Delta\phi$), the length of the basin margin (L), and the fraction of aquifer utilized along the basin margin, F. Focusing might be produced by folds or other structures on the basin margin. An F value of 0.5 would mean focusing and an increase in the escape flow velocity by a factor of two, for example. Substituting the values determined in Figure 10 into the expression in Figure 11, we find that a typical expulsive fluid pulse can carry the temperature of the deep parts of the basin to the near-surface along a basal aquifer, provided the basal aquifer is less than 30/F meters thick. More importantly, perhaps, the expression indicates that basins favorable for Mississippi Valley-type mineralization should be wide, steep sided, have a thin basal aquifer, and have structures that tend to channel flow into or along only certain portions of the basin margin portion of the basal aquifer. That is, W should be large, L small, and F less than one.

Discussion

The above description of basin dewatering deliberately circumvents many difficult details. For example, the porosity of at least some geopressured zones (Jones, 1975) exceeds by much more than a few tenths of a percent the porosity of the surrounding, nongeopressured strata or the porosity that would be anticipated on the basis of the porosity as a function of depth relation and the observed porosity above the geopressured zone. The change in porosity accompanying dewatering might be much larger than the few tenths of a percent used in our calculations. If so, smaller geopressured zones could provide the size fluid pulse required to vent warm fluids to the surface. It is quite possible that, especially as fluid pressures

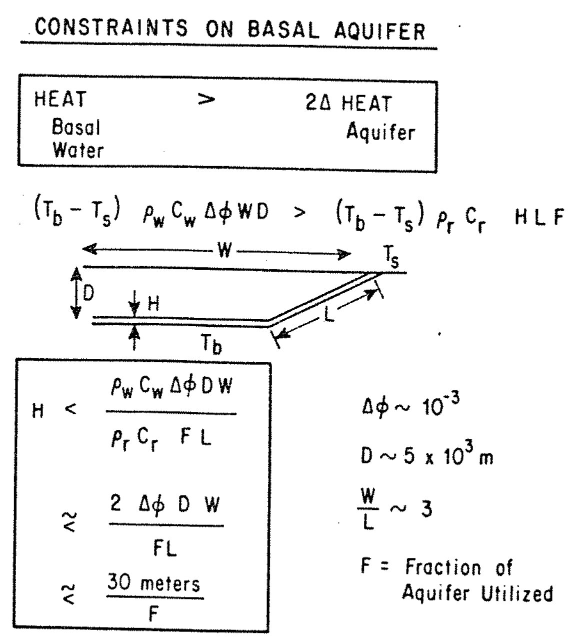


FIG. 11. A minimum requirement for the venting of hot fluids is that the amount of fluid expelled in a single pulse be able to heat the basal aquifer.

approach lithostatic values, geopressured zones will leak. Such leakage would delay and perhaps eliminate rupture. Some leakage will certainly take place through the process of shale membrane filtration (Graf, 1982). As Graf points out, shale membrane filtration would be a very good way to increase the salinity of the pore fluids that remained behind, since the fluid squeezed through the shale membrane would be fresh and thus leave increasingly saline fluids behind. Filtration will certainly occur in geopressured shales and may be a good method for generating the brines that deposited the metals in Mississippi Valley-type deposits.

We have not discussed at all how a geopressured zone in a shale would vent. For venting to be possible, a house-of-cards principal must apply. Collapse of one part of the geopressured zone must put additional stress on adjacent portions and trigger their collapse. During the brief period of collapse and expulsion, the permeability of the geopressured zone must be much greater than it was in the geopressured state (see Sharp, 1978, and Introduction to this paper). The collapse process may be related to the formation of slaty cleavage (Tullis, 1976). House-of-cards-type collapse is not physically implausible; we make no effort in this paper, however, to argue that it is reasonable. Our argument is simply that some process like this must be operative if Mississippi Valley-type deposits are to form as the result of episodic fluid expulsion from basins.

We also have not discussed how fluids from a ruptured geopressured zone could migrate down to a basal aquifer. The increase in lithostatic pressure with depth makes it quite likely that fluids will be expelled upward from a geopressured zone. However, a downward increase in permeability with depth will tend to turn fluids venting upward from a geopressured zone around. If the geopressured zone was somewhat sausage- or fingerlike in geometry—that is, if fluid flow around the sides of the geopressured zone were possible—the released fluids could move into a basal aquifer.

The effect of the dewatering of one geopressured pod on the dewatering of other neighbor pods will also eventually have to be considered.

The above complexities are recognized but have been deliberately sidestepped. It will clearly take a long time to address them all satisfactorily. An experimental approach would perhaps be most profitable for many. It is encouraging to note, however, how little of a basin's fluid is actually required to precipitate the total estimated Mississippi Valley mineralization associated with a basin. For example, the total metal tonnage that might be associated with the Illinois basin (including the Upper Mississippi Valley, Southeast Missouri, Kentucky-Illinois, and Central Illinois districts; Fig. 3) is roughly 100 million tons of combined lead and zinc.[1] The basin contains over 3.8×10^5 km^3 of Paleozoic sediment. Thus, if the ore brines contained 100 ppm Pb plus Zn, a porosity decrease of 0.26 percent could provide the required ore fluid. It may be that economically significant lead-zinc accumulations require extraction of components from a smaller fraction of a basin's pore fluid than would be required for significant oil accumulations.

Other lines of reasoning indicate that the magnitude and rates of episodic fluid expulsion estimated in the previous section may be of the right magnitude. The 50 episodes of fluid expulsion estimated in Figure 9 to occur between a 3- and 5-km depth could account for the sphalerite banding observed in the Upper Mississippi district noted by McLimans et al. (1980). McLimans et al. found about 30 major dark bands interspersed with light bands. Each dark band might represent an episode of geopressure release. If so, the time duration between each band might be about a million years, and the time required to produce all the bands would be about 30 m.y. (see Fig. 10 and previous discussion). In this view the dark portions of the bands would be the central, hottest portions of the pulse, and would carry the most iron. The white bands would represent the cooler leading and trailing edges of the pulse. For the great majority of the time, when expulsed basin fluids were not coursing through the deposits, the deposition sites, being near-surface, would likely be bathed in relatively fresh and much cooler ground water. At these times the sphalerite bands could be dissolved. Dissolution zones should thus be found in the middle of white bands. A similar story could be made for at least eight pulses of mineralization interspersed with periods of dissolution noted by Sverjensky (1981) in the Buick mine of the Viburnum Trend. The episodic pore fluid expulsion model thus provides a convenient explanation for the ore banding and pulses of mineralization interspersed with periods of ore dissolution observed in Mississippi Valley-type deposits. Convincing alternate explanations are difficult to find.

Pulses of fluid expulsion, each from a slightly different part or mixture of parts of the basin, could also account for the geochemical and isotopic complexity of the ore deposits (Sverjensky, 1981; Heyl et al., 1974). Pulses of geopressured fluid, passing through the ore deposition sites, could account for the tectonic features of the deposits and the intimate relationship of these features to ore, as noted by Dozy (1970) and Ohle (1980) and discussed in the Introduction of this paper.

Several estimates of the time that warm mineralizing fluids were present in the sites of ore deposition

[1] McLimans, 1977; Morris et al., 1973; Wedlow et al., 1973.

TABLE 2. Width of Thermal Halos Associated with Escape Aquiters

t	d	F
10 yrs	35 m	0.46
100 yrs	112 m	0.21
1,000 yrs	350 m	0.08
10,000 yrs	1,120 m	0.03

If fluid flows through an aquifer and keeps it anomously warm for a period of time, t, impermeable strata on both sides of the aquifer will be heated; $d = 2\sqrt{Kt}$ approximately describes the equivalent warming of both sides of the aquifer (see Carslaw and Jaeger, 1959, p. 61). Taking the thermal diffusivity, K, to be 0.01 cm^2/sec and t as indicated in the table, we calculate the d values listed. The degree of channeling, F, in the basin margin portions of the aquifer required to preserve the original thermal mass of the 30-m-thick aquifer is also shown. An upper limit on the duration of a thermal pulse is given by the relations in Figure 7. If mineralization occurs over a 50-m.y. period, and dewatering takes place less than 1/5,000th of the time, the maximum duration of a dewatering pulse is 10,000 years. Thus, the bottom line in the table is approximately the maximum duration of a dewatering pulse that would be useful for depositing lead-zinc mineralization

are comparable. In our model we seek to warm only the basal aquifer and as little of the adjacent less permeable formation as possible. Table 2 estimates the thickness of adjacent formation that will be heated by fluid pulses of various durations. The table shows that as the duration of the pulse is increased, greater thicknesses of adjacent formation are heated and a greater focusing of flow through the aquifer at the margins of the basin is required to meet the criteria set forth in Figure 11. The table suggests that the most likely duration of a single expulsive pulse is a few hundred to a few thousand years.

Assuming the ore deposition phase of a basin lasts no longer than 50 m.y., Figure 7 places an upper bound of 50,000 years for the duration of a single fluid pulse (since the rate of fluid expulsion from the basins must be $\gtrsim$1,000 times normal).

Lavery and Barnes (1971) have estimated the total time that mineralizing solutions were present in veins in the Wisconsin-Illinois district from diffusion halos of zinc observed around the veins. From one vein they estimate 2,000 years for the duration of the mineralizing event, from another 250,000 years. Under the assumptions made in Lavery and Barnes' calculations, the mineralization could equally well be a single event or a series of separate events. If mineralization took place in 30 pulses, the time duration of each pulse would be 66 to 8,300 years, estimates which overlap our range of estimates.

Finally A. Gize (pers. commun.) has examined the maturity of hydrocarbons near the Wisconsin-Illinois district and shows that they could not have been exposed to the temperatures of ore deposition for more than 1 m.y. If ore deposition occurred in 30 pulses in this area, the duration of a single pulse would have been 33,000 years.

Various estimates of the time duration of mineralizing pulses thus give similar results. It is reasonable to suppose that if basins are to spawn lead-zinc mineralization, they will do so when they are deep enough to have basal temperatures in excess of 100°C and are subsiding, accumulating sediments, and compacting fairly vigorously. The duration of this period is ~30 m.y. for most basins. The estimates of our heuristic model thus are fairly robust. From several points of view it seems reasonable that Mississippi Valley-type mineralization was deposited over a period of a few tens of millions of years but that the duration of ore deposition was a few thousandths or less of this total period.

Finally, although bursting of a geopressured zone is rare (once every million years or so), it is not inconceivable that the phenomena could be observed somewhere in the world in the next decades. Surface venting of warm solutions might lead to slumping, death of vegetation, and hydrothermal alteration.

Comparison of the Characteristics of Basins with and without Abundant Lead-Zinc Mineralization to the Characteristics Considered Favorable to Lead-Zinc Mineralization under the Episodic Pore Fluid Expulsion Hypothesis

The episodic pore fluid expulsion hypothesis discussed above identifies basin characteristics favorable to the formation of lead-zinc deposits:

1. The basin should contain abundant units with very low permeability (e.g., shale). Rock units with particularly low permeability ($\lesssim$0.005 mD for sedimentation rates of 25 m/m.y.) are required for the formation of geopressured zones (e.g., Sharp, 1978). Lower sedimentation rates require lower basin permeabilities.

2. The internal structure of the impermeable units should be complex, and average regional permeability should increase downward overall. Combined, this will facilitate downward movement of fluids to a basal aquifer.

3. The basin must have a stable margin. One fluid pulse will not deposit enough ore to make an ore deposit. The metal content of many pulses must be deposited in the same near-surface locality if an ore deposit is to be formed.

4. The basin should have a thin but high permeability basal aquifer, and it is preferable that the margins of the basin be steep. If this is the case, a

smaller fluid pulse would push hot deep basin brines to the near-surface without complete cooling.

5. The basin should have structures that focus dewatering fluids out to restricted portions of the margin. Focusing of the flow will allow smaller expulsive pulses to be effective and will in addition produce larger ore deposits by channeling outflow into only a few major escape localities.

One way to test the episodic pore fluid expulsion model is to examine several basins to see if their propensity toward major accumulations of lead-zinc mineralization can be explained in terms of the above criteria. We will briefly examine six basins in this fashion. Three of the basins, the Illinois' Arkoma-Ouachita, and the Appalachian, are geographically associated with known, economically exploitable Mississippi Valley-type lead-zinc deposits; the other three, Michigan, Forest City, and Anadarko, have no known economic Mississippi Valley-type deposits associated with them. Table 3 and the following discussion summarizes some of the salient characteristics of these basins and their associated ore deposits.

Basins with Pb-Zn mineralization

The Illinois basin (Fig. 1) is surrounded by at least five major Pb-Zn districts: Upper Mississippi Valley, Viburnum Trend, Illinois-Kentucky, Central Kentucky, and the Central Missouri districts. Many minor occurrences and districts are also known to exist around the basin margins.

The Illinois basin started out as a broad depression covered by thin but extensive basal sand units. With time the basin developed arches capped by reefs that could host Mississippi Valley-type deposits. Some of these arches were persistent features, such as the Precambrian knobs of the Viburnum Trend that have continuously persisted since the Cambrian. Others were ephemeral. The arches that define the present-day Illinois basin were established mainly in Late Ordovician time.

The thin basal sand units are ideal aquifers for mineralizing solutions. In the Wisconsin-Illinois district, for example, mineralization occurs in the Ordovician Trenton Group (McLimans, 1977), which overlies the permeable St. Peter Sandstone. The St. Peter Sandstone is variable in thickness, averaging 30 m across northern Illinois (Willman et al., 1975), and becomes sandier to the north from the central portion of the basin (Dapples, 1955). It overlies a major Lower Ordovician unconformity that cuts progressively into the section to the north (Willman et al., 1975). Therefore, all fluids flowing to the north edge of the basin would drain into the St. Peter Sandstone.

Mineralization in the Viburnum Trend occurs primarily in the Bonneterre Dolomite, which is capped by the impermeable and sometimes mineralized Davis Shale. The basal unit in the area is the Lamotte Sandstone which lies unconformably on the Precambrian basement and thins and/or pinches out over the granite knobs of the Ozark uplift.

The Illinois-Kentucky and Central Kentucky districts are located over the deeper parts of the Illinois basin. Fluid exodus is thought to have been controlled here by deep vertical faults that have surface expression and penetrate to basement. These faults became active during the Carboniferous. Mineralization in both districts occurs primarily in Mississippian rocks.

The Illinois basin contains over 3.8×10^5 km^3 (after Swann and Bell, 1958) of sediment of which over 20 percent is estimated to be shale. Over 5 km of sediment was deposited in the deepest parts of the basin in the Paleozoic at an average rate of about 18 m/m.y. The pre-Carboniferous shales are fairly continuous throughout the basin and are not as structurally complex as the Carboniferous shales. The Carboniferous shales were deposited more rapidly. Two kilometers of Carboniferous shale were deposited at a rate of approximately 30 m/m.y. These more rapidly deposited shales are intermixed with discontinuous sandstone units. The Carboniferous strata are less permeable than the pre-Carboniferous strata below.

Locally complex folds and major faults occur in the interior of the Illinois basin that could intercept and focus ore-bearing fluids as they migrated to the basin margins.

The Illinois basin thus has many of the characteristics that favor the formation of Mississippi Valley-type mineralization according to the pore fluid expulsion model, including shale with structural complexities, relatively stable positive margins, thin basal aquifers, and structural features that could focus fluid flow to particular parts of the basin margin.

The tectonically and geographically related Arkoma and Ouachita basins (Viele, 1973; Morris, 1975) formed in Carboniferous time as the result of collision of South and North America and are thus termed foreland basins. The collision zone to the south was of course not a stable margin during basin formation (it lies within the so-called Ouachita mobile belt; Viele, 1973). The eastern and western margins were not well defined either, but the northern margin (the Ozark uplift) was stable throughout the Paleozoic, and it is on this margin that the Tri-State and northern Arkansas lead-zinc districts are located. Mineralized fault breccias and shales of Pennsylvanian age indicate that mineralization took place in both districts during Upper or post-Pennsylvanian time (McKnight and Fisher, 1970; Leach et al., 1975). Host rocks for

TABLE 3. Basin Characteristics

Basin	Associated districts	Banding[1]	Potential regional aquifers	Age of host rock	Volume of sediment (km^3)	% shale (estimate)
I. Illinois	A. Upper Mississippi Valley	1, 2	St. Peter Sandstone	Ordovician	>3.8 × 10^5	>20 (?)
	B. Viburnum Trend	1?, 2	Lamotte Sandstone	Cambrian		
	C. Illinois-Kentucky	1, 2	Faults	Mississippian		
	D. Central Kentucky	1, 2	Faults	Mississippian		
	E. Central Missouri	3	Lamotte Sandstone or unconformity	Ordovician → Pennsylvanian		
II. Arkoma-Ouachita	A. Tri-State	2?	Simpson Group, faults	Mississippian	>2.3 × 10^5	>20 (?)
	B. northern Arkansas	3	Simpson Group, faults	Ordovician and Mississippian		
III. Appalachian	A. East Tennessee	1, 2	Cambrian quartzite, limestone, dolomite	Lower Cambrian, Ordovician	>2.15 × 10^6	35
	B. Friedensville	1, 2	Hardyston quartzide (?)	Lower Ordovician		
	C. Central Tennessee	1, 2	Knox dolomite	Ordovician		
IV. Michigan	A. minor	3			>3.8 × 10^5	18
V. Forest City	A. minor	3			>3.3 × 10^5	?
VI. Anadarko	A. western Kansas (?)	3	Simpson Group (?)		>4.13 × 10^5	>20
	B. minor	3				

[1] 1 = within minerals (e.g., sphalerite), 2 = by minerals (paragenesis), 3 = uncertain
[2] P = poddy, discontinuous; c = continuous; ( ) = probable
Characteristics of three basins with associated Pb-Zn districts are compared with those of three basins with no known Pb-Zn districts; the volume of sediment and rates of sedimentation are minimums based on calculations of preserved thicknesses

the ore are Mississippian carbonates in the Tri-State district (Hagni, 1976) and are Mississippian and Ordovician carbonates in North Arkansas.

The lower Paleozoic sequence contains several sandstones and dolomite units of the Simpson Group, which thin over the Precambrian highs in the north and could have acted as the basal aquifers for the mineralizing fluids in both districts. The permeability of the Simpson Group sands tends to increase to the north toward the Ozark uplift (Dapples, 1955).

The Arkoma-Ouachita basin is thickest to the south in the Ouachitas (more than 6 km deep) and thins rapidly to the north of the Arkoma basin over the Ozark uplift. The basin contains well over 2.3 × 10^5 km^3 of sediment (after Huffman, 1959) of which over 20 percent is shale. The major pulse of clastic sedimentation occurred during the Carboniferous in the Arkoma basin. Sedimentation rates ranged from 80 to 60 m/m.y. during these pulses compared with an average rate of 18 to 22 m/m.y. during the entire Paleozoic (Table 3). The shales rapidly deposited during the Carboniferous are complex and poddy compared with the areally continuous and thin shales of the Devonian and older rocks (Amsden, 1981; Viele, 1973; Morris, 1975). They were deposited during the northward migration of the Ouachita orogenic belt. The carbonates and sandstones that were deposited more slowly in the lower Paleozoic may be more permeable regionally than the overlying complex Carboniferous shales and flysch.

The Appalachian basin (Fig. 1) is another example of a fore-arc basin. The Friedensville deposit, the East Tennessee district, and possibly the Central Tennessee district are associated with the Appalachian basin. Host rocks for the deposits range in age from Lower Cambrian to Middle Ordovician and are dolomitized limestones, reefs, and shales.

Potential thin basal aquifers that underlie each of the major districts include the Hardyston Quartzite in eastern Pennsylvania, Lower Cambrian quartzites

TABLE 3. —(*Continued*)

	Major periods of clastic sedimentation					Paleozoic (340-m.y.) sedimentation	
Location	Age	Duration (m.y.)	Complexity of shales[2]	Subsidence rate (m/m.y.)	Thickness of sediment (km)	Subsidence rate (m/m.y.)	Max. thickness of sediment (km)
Basinwide	Mississippian-Pennsylvanian	65	P	30	~2	18	5
Arkoma		35	P	60	>2.1	>18	>5
Ouachita	Mississippian-Pennsylvanian	65	P	80	>5.2	>22	>6.1
Eastern Pennsylvania	Upper Ordovician-Lower Silurian	~15	P, C	~290	~4.4	40	11
	Upper Devonian	~25	P, C	~190	~4.8		
	Mississippian-Pennsylvanian	~65	P	45	~2.9		
East-Central Tennessee	Upper Ordovician-Lower Silurian	~15	P, C	~50	~0.7	17	5
	Upper Devonian	~25	P, C	~30	~0.8		
	Mississippian-Pennsylvanian	~65	P, C	~15	~1.0		
Basin center	Mississippian-Pennsylvanian	65	C, (P)	~8	0.5	14	4
Basin center		35	(P)	24	0.85	8	2.2
Basin wide		35	P	160	5.5	>30	>10.2

References for indicated parts of the table are:

I. A. Hall and Friedman, 1963; Heyl, 1968; Pickney and Rafter, 1972; McLimans, 1977; McLimans et al., 1980. B. Roedder, 1977; Snyder and Gerdemann, 1968; *Economic Geology*, 1977; Sverjensky, 1981. C. Roedder, 1967, 1976; Grogan and Bradbury, 1968; Heyl et al., 1966. D. Jolly and Heyl, 1964; Snyder, 1968; Plummer, 1971; Roedder, 1971, 1976; could be associated with Appalachian basin. E. Leach, 1980; may be associated with Arkoma basin.

II. A. Hagni and Grawe, 1964; Brockie et al., 1968; Hagni, 1976. B. Snyder, 1968; Leach et al., 1975

III. A. Kendall, 1960; Heyl et al., 1966; Crawford and Hoagland, 1968; Hoagland, 1976. B. Callahan, 1968; Roedder, 1971, 1976. C. Snyder, 1968; Roedder, 1971, 1976; Kyle, 1976; Hoagland, 1976.

and carbonates in East Tennessee, and the Knox Dolomite in Central Tennessee. Faults, folds, and unconformities could have acted as focusing mechanisms for the ore-forming solutions.

The Appalachian basin contains over 2.15×10^6 km^3 of sediment, which is estimated to be at least 35 percent shale (Colton, 1970). Today the basin is thickest in the east (greater than 5 km) and thins rapidly to the west over a stable hinterland. Unlike the basins discussed previously, however, the Appalachian basin was influenced by at least four major orogenic events: the Grenville, Taconic, Acadian, and Alleghenian orogenies. During each, the rate of sedimentation increased dramatically (Table 3). The shales associated with each pulse of sedimentation are structurally complex. The shales deposited during times of less rapid sedimentation are fairly continuous, often thin, and interbedded with relatively continuous sandstones and carbonates.

The unknown age of the ore deposits is a major barrier to a more detailed understanding of their origin. For example, there are more than 1.5×10^5 km^3 of Ordovician shales in the central parts of the Appalachian basin that are younger than the strata that host the ore deposits (Hoagland, 1976; Colton, 1970). These shales are part of the flysch sequence deposited during the Taconic orogeny. Sediments reached a thickness of over 7 km in parts of the basin at the end of the Taconic orogeny. Could these shales have been involved in the mineralization? We simply don't know because we don't know the age of the deposits.

Both the Arkoma-Ouachita and Appalachian basins thus have substantial accumulations of low permeability strata with complex structure overlying more permeable carbonates and sandstones. Both basins have structures that could focus discharge and the margins that host ore were stable at the time of mineralization. Like the Illinois basin, these basins appear to meet the criteria for the basin expulsion model and all three basins indeed have associated Mississippi Valley-type ore deposits.

Basins with no major Pb-Zn districts around the basin margins

The Michigan basin (Fig. 1) is a circular, intracratonic basin like the Illinois basin, surrounded by major tectonic arches that were positive features in the lower Paleozoic (Cohee, 1965). The basin contains over 3.8×10^5 km^3 of sediment (18% shale, Colton, 1970) and reached a thickness of over 3 km during the Devonian. The history of lower Paleozoic sedimentation (but not sediment type) and tectonics is similar to the Illinois basin (Hamblin, 1958; Melhorn, 1958; Cohee, 1948; Fowler and Kuenzi, 1978). Carbonates and evaporites are the dominant lithologies with only minor shales. The evaporites could be a source of the high salinity brines. The Mt. Simon and St. Peter Sandstones are possible aquifers for ore-forming solutions. Potential reef and dolomite host rocks are numerous and occur throughout the Paleozoic. However, unlike the Illinois basin, there was never a major pulse of clastic sedimentation with its associated complex and poddy shales. Instead, the major shales in the Michigan basin are relatively thin, extensive, and continuous. The shale geometry would not favor the downward migration of expulsed fluids.

The Michigan basin is also characterized by very little internal structure. Few anticlines and deep basin faults are known. There appears to be no major mechanism capable of focusing ore solutions to one part of the basin margin.

The Forest City basin (Fig. 1) is another intracratonic basin that contains more than 3.3×10^5 km^3 of sediment. The basin, however, never reaches a thickness of more than about 2.2 km of preserved sediment and thus may have been relatively cool. Provided suitable temperatures for Mississippi Valley-type mineralization were attained (and they could well have been), Mississippi Valley-type deposits could have been produced, because the basin's tectonic and sedimentologic history is similar to that of the Illinois basin, and the general condition for deposit formation is favorable. Potential host rocks, basal aquifers, complex shales, and internal basin structure, folds, and faults that could have focused fluid outflow are present in the Forest City basin.

The Anadarko basin (Fig. 1) is a fore-arc basin similar to the Arkoma-Ouachita basin. It contains over 4.5×10^6 km^3 of Paleozoic sediment, more than 20 percent of which is shale. Over 5.5 km of the more than 10 km of sediment deposited in the deep parts of the basin were deposited in the Pennsylvanian at a rate of over 160 m/m.y. The basin is thickest in the south and thins gradually to the north over a stable hinterland west of the Ozark uplift. Major tectonic structures in the area are Carboniferous or younger. There are no major arches or highlands in the stable margin area over which the pre-Carboniferous sediments thinned. Potential fluid-focusing structures, thrusts, faults, folds, and basal aquifers are present. The thinning of the basin to the north seems to have simply been too gradual to allow venting of the warm basal fluid required for mineralization.

The basins without economic lead-zinc deposits thus lack one or more of the five criteria required to form the deposits by the fluid expulsion model. Either they do not have complex impermeable strata (Michigan basin), they were not deep enough to develop warm enough basal brines (Forest City basin), they lacked focusing structure in the basal aquifers (Michigan basin), or they lacked sufficiently steep or stable margins (Anadarko basin). The criteria for formation of Mississippi Valley-type mineral deposits deduced from our episodic pore fluid expulsion model appear to have some utility in distinguishing basins favorable for mineralization from those that are unfavorable.

Conclusions and Summary

We have examined in some detail one aspect of the basinal brine-pore fluid expulsion hypothesis for formation of Mississippi Valley-type lead-zinc deposits—namely, the rates of fluid flow up a basal aquifer that would be required to carry, in accordance with this hypothesis, the temperatures of the deep parts of the basin to near-surface sites of ore deposition. We have shown that the flow rates required for most basins are >300 to 5,000 times those that could be produced by normal steady basin subsidence, sedimentation, and compaction (see Fig. 7). If the pore fluid expulsion hypothesis is to work, it appears the expulsion of pore fluids from a basin must be an episodic phenomenon. Pore fluids must be expulsed at thousands of times the average rate over periods of time representing only a few thousandths of the pertinent history of basin growth.

The implications of the episodic pore fluid expulsion hypothesis for lead-zinc deposition have been explored in a first cut fashion. If dewatering occurs in a cycle as fluid pressure builds up to lithostatic levels, approximately 50 dewatering pulses might occur, one every million years or so, as a strata subsides from 3 to 5 km in depth (Fig. 10). Roughly this number of dewatering episodes could account for the major color bands in sphalerite from the Upper Mississippi Valley district and for the eight or so pulses of mineralization observed in mines of the Viburnum

Trend. In between dewatering pulses the near-surface sites of ore deposition would most likely be exposed to surface waters that might corrode the sulfide minerals. Thus, the observed cycles of deposition and corrosion can be explained by episodic dewatering. As discussed by Dozy (1970) and reviewed in the Introduction, many of the tectonic features of Mississippi Valley-type deposits and the close relation of these features to ore deposition (Ohle, 1980) are naturally explained by the periodic passage of pulses of geopressured fluids. The same general number and duration of dewatering pulses are indicated by several independent lines of evidence.

Using the fact hot basal brines must at least heat up the hypothetical escape aquifer at the margin of the basin if hot fluids are to reach the near-surface sites of ore deposition, it is shown that ore deposition should be favored by wide basins with steep sides, a thin basal aquifer, and structures that channel flow to restricted portions of the basin margin (Fig. 11). Impermeable strata are of course required if geopressured zones are to develop. This means a basin favorable for lead-zinc mineralization should have abundant shale or flysch. Complex sedimentology or structure and permeability that increases downward will promote downward migration of expelled pore fluids and these should, therefore, also be considered favorable features. Because one fluid pulse is not sufficient to make an ore deposit, and tens of millions of years are probably required to accumulate enough pulses, favorable basins must have stable margins. The sites of ore deposition (fluid escape) must remain near the surface for prolonged periods, if economically interesting accumulations of metals are to be deposited.

Five basins in the mid-continent region of the United States were reviewed in light of the above criteria for Mississippi Valley-type mineralization potential. Within our limited and perhaps biased knowledge of the basins, the above criteria appear to be successful for distinguishing basins with superior lead-zinc mineralization potential (as evidenced by presently known ore deposits) from basins with little or no potential. The criteria set forth above and discussed in the body of this paper may thus have some exploration utility.

The low permeability ($\sim$0.005 mD) required to develop geopressured zones in a basin subsiding at a few tens of meters per million years is low enough to preclude any significant migration of hydrocarbons or free convection of water (Cathles, 1981). Hydrocarbon migration, to the extent it does occur in basins that produce copious Mississippi Valley-type mineral deposits, thus probably occurs during rupture of geopressured zones and episodic pore fluid release. Study of lead-zinc mineralization associated with basins, and particularly the banding of such mineralization, might thus make a contribution to understanding hydrocarbon migration in these basins. The episodic movement of pore fluids we envision is certainly different from the steady upward movement envisioned by Bonham (1980) to occur predominantly in more permeable basins with significant oil reservoirs. The fact the permeability required by the episodic expulsion hypothesis is low enough to preclude free convection means that this hypothesis is cleanly separated from any hypothesis that involves fluid circulation around an igneous intrusive.

The episodic dewatering hypothesis appears to contain all the ingredients necessary to produce the lead-zinc mineralization associated with basins. The fluid that will be eventually expelled must spend long periods of time in a geopressured state enclosed by impermeable strata. Under such conditions reverse osmosis will operate and the fluids in the geopressured zone will become more saline. Sudden rupture could expel the fluids at the rates required to carry temperatures typical of the deep portions of the basin to the near-surface. Sverjensky (1981) has shown the fluids could carry tens of parts per million of both lead and sulfide sulfur in the ratio stoichiometrically required to form galena. Mineral deposition could occur near the surface due to simple cooling.

The most important conclusion of this paper is that episodic dewatering is required if temperature gradients at the basin margins are to be perturbed and the conditions for lead-zinc mineral deposition established. Beyond this the rough outlines of an episodic dewatering model for Mississippi Valley-type mineralization have been sketched. We hope this may serve as a useful framework for future studies.

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REFERENCES

Amsden, T. W., 1981, Arkoma basin model: Middle Ordovician through Early Devonian [abs.]: Am. Assoc. Petroleum Geologists Bull., v. 65, p. 1495.

Anderson, G. M., and MacQueen, R. U., 1982, Ore deposits models-6. Mississippi Valley lead-zinc deposits: Geosci. Canada, v. 9, p. 108–117.

Beales, F. W., and Jackson, S. A., 1966, Precipitation of lead-zinc ores in carbonate reservoirs as illustrated by Pine Point ore field, Canada: Inst. Mining Metallurgy Trans., v. 75, sec. B, p. 278–285.

Beales, F. W., and Onasick, E. P., 1970, The stratagraphic habit of Mississippi Valley-type ore bodies: Inst. Mining Metallurgy Trans., v. 79, sec. B, p. B145–B154.

Boles, J. R., 1981, Clay diagenesis and effects on sandstones cementation (case histories from the Gulf Coast Tertiary), *in* Longstaffe, F. J., ed., Short course in clays and the resource geologist: Calgary, Mining Assoc. Canada, May 1981, p. 148–168.

Bonham, L. C., 1980, Migration of hydrocarbons in compacting basins: Tulsa, Oklahoma, Am. Assoc. Petroleum Geologists, Studies in Geology 10, p. 69–88.

Brockie, D. C., Hare, E. H., Jr., and Dingess, P. R., 1968, The geology and ore deposits of the Tri-State district of Missouri, Kansas and Oklahoma, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Inst. Mining Metall. Petroleum Engineers, v. 1, p. 400–430.

Callahan, W. H., 1968, Geology of the Friedensville zinc mine, Lehigh County, Pennsylvania, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Inst. Mining Metall. Petroleum Engineers, v. 1, p. 95–107.

Carslaw, H. S., and Jaeger, J. C., 1959, Conduction of heat in solids, 2nd ed.: Oxford, Oxford Univ. Press, p. 101.

Cathles, L. M., 1977, An analysis of the cooling of intrusives by ground-water convection which includes boiling: ECON. GEOL., v. 72, p. 804–826.

—— 1981, Fluid flow and genesis of hydrothermal ore deposits: ECON. GEOL., 75TH ANN. VOL., p. 424–457.

Clark, S. P., ed., 1966, Handbook of physical constants, rev. ed.: Geol. Soc. America, Mem. 97, 587 p.

Cohee, G. V., 1948, Cambrian and Ordovician rocks in the Michigan basin and adjoining areas: Am. Assoc. Petroleum Geologists Bull., v. 32, p. 1417–1448.

—— 1965, Geologic history of the Michigan basin: Washington Acad. Sci. Jour., v. 55, p. 211–223.

Colton, G. W., 1970, The Appalachian basin—its depositional sequences and their geologic relationships, *in* Fisher, G. W., Pettijohn, F. J., Reed, J. C., Jr., and Weaver, K. N., eds., Studies of Appalachian geology: Central and southern: New York, Wiley Interscі. Pub., p. 5–47.

Crawford, J., and Hoagland, A. D., 1968, The Mascot-Jefferson City zinc district, Tennessee, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Inst. Mining Metall. Petroleum Engineers, v. 1, p. 242–256.

Dapples, E. C., 1955, General lithofacies relationships of St. Peter Sandstone and Simpson Group: Am. Assoc. Petroleum Geologists Bull., v. 39, p. 444–467.

Dozy, J. J., 1970, A geologic model for the genesis of the lead-zinc ores of the Mississippi Valley, U.S.A.: Inst. Mining Metallurgy Trans., v. 79, sec. B, p. 163–170.

Economic Geology, 1977, An issue devoted to the Virburnum Trend, Southeast Missouri: v. 72, p. 337–490.

Fowler, J. H., and Kuenzi, W. D., 1978, Keweenawan turbidites in Michigan (deep borehole red beds): A foundered basin sequence developed during evolution of a prototectonic rift system: Jour. Geophys. Research, v. 83, no. B12, p. 5833–5843.

Graf, D. L., 1982, Chemical osmosis, reverse chemical osmosis and the origin of subsurface brines: Geochim. et Cosmochim. Acta.

Grogan, R. M., and Bradbury, J. C., 1968, Fluorite-zinc-lead deposits of the Illinois-Kentucky mining district, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967, (Graton-Sales vol.): New York, Am. Inst. Mining Metall. Petroleum Engineers, v. 1, p. 370–399.

Hagni, R. D., 1976, Tri-state ore deposits: The character of their host rocks and their genesis, *in* Wolf, K. H., ed., Handbook of strata-bound and stratiform ore deposits: New York, Elsevier, v. 6, p. 457–494.

Hagni, R. D., and Grawe, O. R., 1964, Mineral paragenesis in the Tri-State district, Missouri, Kansas, Oklahoma: ECON. GEOL., v. 59, p. 449–457.

Hall, W. E., and Friedman, I., 1963, Composition of fluid inclusions, Cave-in-Rock fluorite district, Illinois, and Upper Mississippi Valley zinc-lead district: ECON. GEOL., v. 58, p. 886–911.

Hamblin, W. K., 1958, The Cambrian sandstone of northern Michigan: Michigan Geol. Survey Pub. 51, 146 p.

Hanor, J. S., 1979, The sedimentary genesis of hydrothermal fluids, *in* Barnes, H. L., ed., Geochemistry of hydrothermal ore deposits, 2nd ed.: New York, John Wiley and Sons, p. 137–172.

Helgeson, H. C., 1969, Thermodynamics of hydrothermal systems at elevated temperatures and pressures: Am. Jour. Sci., v. 267, p. 729–804.

Heyl, A. V., 1968, The Upper Mississippi Valley base-metal district, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Inst. Mining Metall. Petroleum Engineers, v. 1, p. 431–459.

Heyl, A. V., Delevaux, M. H., Zartman, R. E., and Brock, M. R., 1966, Isotopic study of galenas from the Upper Mississippi Valley, the Illinois-Kentucky, and some Appalachian Valley mineral districts: ECON. GEOL., v. 61, p. 933–961.

Heyl, A. V., Landis, G. P., and Zartman, R. E., 1974, Isotopic evidence for the origin of Mississippi Valley-type mineral deposits: A review: ECON. GEOL., v. 69, p. 992–1006.

Hoagland, A. D., 1976, Appalachian zinc-lead deposits, *in* Wolf, K. H., ed., Handbook of strata-bound and stratiform ore deposits: New York, Elsevier, v. 6, p. 495–534.

Howe, S. S., 1981, Mineralogy, fluid inclusions, and stable isotopes of lead-zinc occurrences in central Pennsylvania: Unpub. M.S. thesis, The Pennsylvania State Univ., 98 p.

Huffman, G. G., 1959, Pre-Desmoinesian isopachous and paleogeologic studies in central mid-continent region: Am. Assoc. Petroleum Geologists Bull., v. 43, p. 2541–2574.

Jackson, S. A., and Beales, F. W., 1967, An aspect of sedimentary basin evolution: The concentration of Mississippi Valley-type ores during late stages of diagenesis: Canadian Petroleum Geology Bull., v. 15, p. 383–433.

Jolly, J. L., and Heyl, A. V., 1964, Mineral paragenesis and zoning in the Central Kentucky mineral district: ECON. GEOL., v. 59, p. 596–624.

Jones, P. H., 1975, Geothermal and hydrocarbon regimes, Northern Gulf of Mexico basin, *in* Dorfman, M. H., and Deller, R. W., eds., Proceedings of the First Geopressured Energy Conference: Univ. Texas, Austin, Center of Energy Studies, June 2–4, p. 15–97.

Kendall, D. L., 1960, Ore deposits and sedimentary features, Jefferson City mine, Tennessee: ECON. GEOL., v. 55, p. 985–1003.

Kinsman, D. J. J., 1975, Rift Valley basins and sedimentation history of training continental margins, *in* Fisher, A. G., and Judson, S., eds., Petroleum and global tectonics: Princeton, Princeton Univ. Press, p. 83–126.

Kyle, J. R., 1976, Brecciation, alteration, and mineralization in the Central Tennessee zinc district: ECON. GEOL., v. 71, p. 892–903.

Lavery, N. G., and Barnes, H. L., 1971, Zinc dispersion in the Wisconsin zinc-lead district: ECON. GEOL., v. 66, p. 226–240.

Leach, D. L., 1980, Nature of mineralizing fluids in the barite deposits of central and southeast Missouri: ECON. GEOL., v. 75, p. 1168–1180.

Leach, D. L., Nelson, R. C., and Williams, D., 1975, Fluid inclusion studies in the northern Arkansas zinc district: ECON. GEOL., v. 70, p 1084–1091.

McKnight, E. T., and Fisher, R. P., 1970, Geology and ore deposits

of the Pitcher field, Oklahoma and Kansas: U. S. Geol. Survey Prof. Paper 588, 165 p.
McLimans, R. K., 1977, Ecological fluid inclusion and stable isotope studies of the Upper Mississippi Valley zinc-lead district, southwestern Wisconsin: Unpub. Ph.D. thesis, Pennsylvania State Univ., 175 p.
McLimans, R. K., Barnes, H. L., and Ohmoto, H., 1980, Sphalerite stratigraphy of the Upper Mississippi Valley zinc-lead district, southwest Wisconsin: ECON. GEOL., v. 75, p. 351–361.
Melhorn, W. N., 1958, Stratigraphic analysis of Silurian rocks in the Michigan basin: Am. Assoc. Petroleum Geologists Bull., v. 42, p. 816–838.
Morris, R. C., 1975, Stratigraphy and structure of part of frontal Ouachita mountains, Arkansas: Am. Assoc. Petroleum Geologists Bull., v. 59, p. 747–765.
Morris, H. T., Heyl, A. V., and Hall, R. B., 1973, Lead: U. S. Geol. Survey Prof. Paper 820, p. 313–332.
Nesbitt, H. W., 1977, Estimation of the thermodynamic properties of Na-Ca and Mg-beidellites: Canadian Mineralogist, v. 15, p. 22–30.
Noble, E. A., 1963, Formation of ore deposits by waters of compaction: ECON. GEOL., v. 58, p. 1145–1156.
Nriagii, J. O., 1975, Thermochemical approximations for clay minerals: Am. Mineralogist, v. 60, p. 834–839.
Ohle, E. L., 1952, Geology of the Hayden Creek lead mine, southeast Missouri: Mining Eng., p. 477–483.
—— 1980, Some considerations in determining the origin of ore deposits of the Mississippi Valley type: Part II: ECON. GEOL., v. 75, p. 161–172.
Pickney, D. M., and Rafter, T. A., 1972, Fractionation of sulfur isotopes during ore deposition in the Upper Mississippi Valley zinc-lead district, Illinois: ECON GEOL., v. 67, p. 315–328.
Plummer, L. M., 1971, Barite deposition in central Kentucky: ECON. GEOL., v. 66, p. 252–258.
Rickard, D. T., Willden, M. Y., Marinder, N. E., and Donnelly, T. H., 1979, Studies on the genesis of the Laisvall sandstone lead-zinc deposit, Sweden: ECON. GEOL., v. 74, p. 1255–1285.
Robie, R. A., Hemingway, B. S., and Fisher, J. R., 1978, Thermodynamic properties of minerals and related substances at 298.15K and 1 bar (10^5 pascals) pressure and at higher temperatures: U. S. Geol. Survey Bull. 1452, 456 p.
Roedder, E., 1967, Environment of deposition of stratiform (Mississippi Valley-type) ore deposits, from studies of fluid inclusions: ECON. GEOL. MON. 3, p. 349–361.
—— 1971, Fluid-inclusion evidence on the environment of formation of mineral deposits of the southern Appalachian Valley: ECON. GEOL., v. 66, p. 777–791.
—— 1976, Fluid inclusion evidence on the genesis of ores in sedimentary and volcanic rocks, *in* Wolf, K. H., ed., Handbook of strata-bound and stratiform ore deposits: New York, Elsevier, v. 2, p. 67–110.
—— 1977, Fluid inclusion studies of ore deposits in the Viburnum Trend, southeast Missouri: ECON. GEOL., v. 72, p. 474–479.
Sclater, J. G., and Christie, P. A. F., 1980, Continental stretching: An explanation of post-mid-Cretaceous subsidence of the Central North Sea Basin: Jour. Geophys. Research, v. 85, no. B7, p. 3711–3739.
Sharp, J. M., Jr., 1978, Energy and momentum transport model of the Ouachita basin and its possible impact on formation of economic mineral deposits: ECON. GEOL., v. 73, p. 1057–1068.
Sleep, N. H., 1976, Platform subsidence mechanisms and "eustatic" sea level changes: Tectonophysics, v. 36, p. 45–56.
Snyder, F. G., 1968, Geology and mineral deposits, midcontinent United States, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Mining Metall. Petroleum Engineers, v. 1, p. 257–286.
Snyder, F. G., and Gerdemann, P. E., 1968, Geology of the Southeast Missouri lead district, *in* Ridge, J. D., ed., Ore deposits of the United States, 1933–1967 (Graton-Sales vol.): New York, Am. Mining Metall. Petroleum Engineers, v. 1, p. 327–358.
Sverjensky, D. A., 1981, The origin of Mississippi Valley-type deposits in the Viburnum Trend, Southeast Missouri: ECON. GEOL., v. 76, p. 1848–1872.
Swann, D. H., and Bell, A. H., 1958, Habitat of oil in the Illinois basin, *in* Weeks, L. G., ed., Habitat of oil: A symposium: Tulsa, Oklahoma, Am. Assoc. Petroleum Geologists, p. 447–472.
Tectonic Map of the United States, 1962, U. S. Geol. Survey and Am. Assoc. Petroleum Geologists.
Touloukian, Y. S., and Ho, C. Y., eds., 1981, Physical properties of rocks and minerals: New York, McGraw-Hill Book Co., v. II-2, 458 p.
Tullis, T. E., 1976, Experiments on the origin of shaley cleavage and schistosity: Geol. Soc. America Bull., v. 87, p. 745–753.
Viele, G. W., 1973, Structure and tectonic history of the Ouachita Mountains, Arkansas, *in* DeJong, K. A., and Scholten, R., eds., Gravity and tectonics: New York, John Wiley and Sons, p. 361–377.
Wedlow, H., Jr., Kulsgaard, T. H., Heyl, A. V., and Hall, R. B., 1973, Zinc: U. S. Geol. Survey Prof. Paper 820, p. 697–711.
Willman, H. B., Atherton, E., Buschbach, T. C., Collinson, C., Frye, J. C., Hopkins, M. E., Lineback, J. A., and Simon, J. A., 1975, Handbook of Illinois stratigraphy: Illinois State Geol. Survey Bull. 95, 261 p.

APPENDIX

Glossary

c_R = heat capacity of the water-saturated formation
g = gravitational constant cm/sec^2
k = permeability in cm^2
K = thermal conductivity in cal/cm-sec-°C
p = fluid pressure in dyns/cm^2
q = Darcy mass flux in g/cm^2-sec
t = time in seconds
T = temperature in °C
ν = fluid viscosity in stokes
ϕ = formation porosity
ρ = fluid density
ρ_R = formation density

An equation that describes the compactive expulsion of fluid from a two-dimensional basin can be easily derived from Darcy's law and the equation of fluid continuity. Darcy's law may be expressed (Cathles, 1981):

$$\underline{q} = \frac{K}{\nu}(\underline{\nabla}\rho - \rho\underline{g}). \tag{A1}$$

The equation of continuity assuming no sources or sinks of fluid in the formation may be written:

$$\frac{\partial(\rho\phi)}{\partial t} + \nabla \cdot q = 0. \tag{A2}$$

If we consider that the fluid density is a constant, but that fluid pressure is perturbed by the compaction process and thus a function of space and time, we may write

$$\rho = \rho_0 \tag{A3}$$

$$p = p_0(z) + p_1(x, y, z, t). \tag{A4}$$

Assuming no flow under unperturbed (zero order) conditions,

$$\nabla p_0 - \rho_0 g_0 = 0. \tag{A5}$$

Substituting (A3) and (A4) into (A1) and using (A5), we find

$$\underline{q} = -\frac{k}{\nu}\underline{\nabla} p_1. \tag{A6}$$

Substituting (A6) into the continuity equation (A2), using (A3), and expanding $\partial\phi/\partial t$ using the chain rule and using $\partial z/\partial t = v_z$, a simple Poisson equation for fluid pressure is found:

$$\underline{\nabla}\cdot\frac{k}{\nu}\underline{\nabla} p_1 - \rho v_z \frac{\partial \phi}{\partial z}. \tag{A7}$$

The heat balance equation simplified from Cathles (1977) is:

$$K\nabla^2 T - \underline{\nabla}\cdot\underline{q}T = \rho_R c_R \frac{\partial T}{\partial t}. \tag{A8}$$

Figure 9 shows the finite difference solution of equations (A6), (A7), and (A8). An alternating direction implicit method of solution was used as described in Cathles (1977).