Chapter 8

A Discussion of Flow Mechanisms Responsible for Alteration and Mineralization in the Cambrian Aquifers of the Ouachita-Arkoma Basin-Ozark System

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ABSTRACT

Cathodoluminescent microstratigraphy in epigenetic dolomite cements correlates over a north-south distance of >275 km across the Ozark Mountains. Trace elements in dolomite show coherent regional variations. Ubiquitous coeval Pb-Zn mineralization contains fluid inclusions with homogenization temperatures >100°C. All suggest the flow of brines (mainly) north from the Arkoma Basin through Cambrian sandstone and carbonate aquifers. The observation that brines still fill Ordovician and Cambrian strata ringing the Ozark Plateau constrains the cumulative flow that has occurred. If the Mid-Continent sediment cover was thick and insulating, brine flow driven by topographic differences in hydrologic head could have been slow enough to avoid salt flushing and still accommodate the fluid inclusion homogenization data. If the cover was thermally conductive and thin, as seems most geologically reasonable, flow at the rates required to explain the homogenization temperatures would have quickly flushed salt from the aquifers in contradiction to present observations. Topographically driven hydrologic flow across the Arkoma basin could not have continued unimpeded for protracted periods. Given the present high permeability of the Pb-Zn deposits, there is no obvious way to limit or pulse cross-basin hydrologic flow. The simplest explanation is that brines were expelled by compaction or gas displacement. High temperature, low salinity fluid inclusions in the Ozark Cambrian aquifers probably represent the incursion of meteoric water into outcrop areas warmed by pulses of brine outflow. Channeling of fluid flow and the relation of alteration and fluid inclusions to flow channels need to be further investigated theoretically and in the field.
INTRODUCTION

The flow of brine north from the Arkoma Basin has produced remarkably extensive and coherent alteration in Upper Cambrian sediments across Arkansas, Missouri, eastern Kansas, and eastern Oklahoma (Fig. 1). A correlative catadolomuminous microstratigraphy in hydrothermal epigenetic dolomite cements over a N-S distance of ~275 km (Figures 1B, Voss and Hagni, 1985; Rowan, 1986; Faye and Bankhead, 1989) in the upper portion of an Fm-Zn mineralization core with deposition of the dolomite cement (Voss, 1989; Covey et al., 1987), and four outcrops of the Mississippi Valley-type (MVT) Pb-Zn mining districts, 3 fluid inclusion homogenization temperatures in four mining districts that range from 60 to 160 °C (Leach and Rowan, 1986; Rowan and Leach, 1989), (4) a northward increase in K/Cl fluid inclusions with constant Na/Cl ratios (Vieets and Leach, 1990), and (5) a regular northward decrease in Fe, Mn, and $\text{Sr}^{2+}/\text{Sr}$, and an increase in the Sr contents of epigenetic dolomite within 1 m of the Lambotte Sandstone contact (Gregg and Shelton, 1989a, b). Carbon and oxygen isotopic ratios in the epigenetic dolomites vary in a fashion consistent with thermal excursion(s) from $50^\circ$ up to $130^\circ$ and then back to $50^\circ$ (Gregg and Shelton, 1989a). The spread in fluid inclusion homogenization temperatures, banding and dissolution stratigraphy, episodes of metal deposition with dissolution in-between (Sverjensky, 1981, Hagni, 1976), and the wide range of lead isotopic ratios in single galena crystals (Hagni, 1976; Hart et al., 1981) indicate variable physical or chemical conditions. Structural control of brine flow from the Arkoma Basin by the Redrock Rift (Farr, 1989) and brine flow within the Ouachita Basin (Gregg and Shelton, 1989b) are suggested by the pattern of trace elements in the dolomite cements. The Ozark-Arkoma system is similar in many respects to other basin margin systems, especially for example the Marathon-Peterson-Mammoth Basin-Llano uplift, the Western Canada Basin, and the Appalachian basin, all of which host MVT lead-zinc deposits.

Two flow mechanisms are generally considered possible explanations for the MVT lead-zinc deposits and their associated epigenetic alteration: Bethke (1986); Bethke et al. (1988); and Meny (1984, 1985) proposed that MVT deposits are produced when basin brines are moved out of a basin by gravity flow from highland areas such as the Ouachita Mountains. This hypothesis is consistent for the Ouachita Mountains (e.g., Leach and Rowan, 1986) because the mineralization and alteration found there require the throughput of large fluid volumes. Given enough volume, this could mobilize limited quantities of fluid. The other flow mechanism that has been suggested for MVT deposits and their associated alteration is the expulsion of brines by compaction or processes related to sediment heating such as gas generation and water displacement from the accumulating and deforming sediment piles (Sharp, 1978; Cathles and Smith, 1983; Oliver, 1986; Rich, 1927). The volume of brine that can be expelled by these mechanisms is limited but could provide an important mechanism for mineralization and alteration. The purpose of this chapter is to examine these competing mechanisms of fluid flow with reference to the alteration and mineralization in the Ouachita-Arkoma Basin-Ozark system.

THE OUACHITA-ARKOMA BASIN-OZARK SYSTEM

The flow system considered in this paper is shown in Figure 1. What is today the Ouachita orogenic belt was a rifted Atlantic-type margin in early and middle Paleozoic time (Housenek, 1886; Arbenz, 1989). The 0–150-m-thick, quartzose Lamotte sandstone was laid down unconformably on Precambrian basement and covered by shallow marine carbonates and shales from Cambrian to Early Pennsylvania time. The maximum thickness of these strata is 1.5 km. Overhanging from the south in Pennsylvania time flexed the margin down to form a classic foreland basin. Sedimentation increased abruptly at ~310 Ma (Arbenz, 1989, figure 7). Five and a half kilometers of Atokan sediments were deposited. Five kilometers of Ordovician strata and the upper portion of the Upper Cambrian are very well mineralized. This is consistent with deformation and sedimentation coalescing after Atokan time.

The Cambrian and Ordovician stratigraphy of the area is shown in Figure 2. From a diagenesis-fluid flow perspective, perhaps the most important aspect of the transect shown in Figure 1 is the microstratigraphy displayed by epigenetic dolomite in the Upper Cambrian to basal Atokan units. The Lamotte Sandstone appears to have been the main conduit for the flow of mineralizing brines, although brines also moved through the overlying Cambrian dolomites and underlying granitic basement. Hydrothermal dolomite strata have a characteristic cathodoluminescent banding. As indicated in Figure 1B, hydrothermal dolomite banding and cement dissolution events can be correlated over ~270 km-long. Voss and Hagni, 1985, between the Viburum Trend and the Northern Arkansas Pb-Zn district (Rowan, 1986), and for 70 km north by north Viburum Trend (Farr, 1989; Voss et al., 1989; Gregg, 1983), a distance of at least 275 km along this section.

The nature of the cathodoluminescent banding varies spatially. The volume of bands increases abruptly near the Bonnette-Lamotte transition and then decrease upward with much fewer and lower intensity bands in shale units (Farr, 1989). These relationships are expected because: (1) iron above a certain concentration inhibits luminescence, (2) Fe and Mn (the main fluorescent agent) are consumed by dolomitization, and (3) shale inhibits fluid throughput and therefore alteration (Farr, 1989). In the middle and lower Bonnette the number of bands and the percentage of duff bands increase from the St. Francois Mountains (which lie between the Viburum Trend (VT) and Old Lead Belt (OLB) districts in Figure 1A) to both the south and east. This is consistent with relatively Fe- and Mn-rich brines reaching the St. Francois Mountain area both by flowing directly north out of the Arkoma and by flowing from the Arkoma northeast along the Redrock Rift and then east along the northwest striking Broomsfield Lineament that connects the Redrock with the St. Francois Mountains area. Deposition of distinct sulfide minerals occurs district wide between particular cathodoluminescent zones (Voss et al., 1989).

An excellent and comprehensive review of all aspects of the chemical and isotopic alteration of the area that supports and augments the inferences drawn from dolomite cathodoluminescence has been given by Gregg and Shelton (1989a). Trace elements in the hydrothermal dolomite within 1 m of the Bonnette-Lamotte transition show an increase in Fe and Mn and a decrease in Sr from the St. Francois Mountains to the south (Gregg and Shelton, 1989b). The pattern is absent 3 to 6 m above the contact, however, indicating less flow in that part of the Bonnette. Similar Se, Cu, and Ag trends from the St. Francois Mountains to the northeast suggests that brines arrived in the same cathodoluminescent time bracket from the north. Flow from the Illinois basin as well as from the Arkoma basin seems to have contributed to alteration and mineralization (Gregg and Shelton, 1989a,b). The decrease in Fe and Mn upward in the Bonnette from the Lamotte sandstone, also as Mn, Fe, Cu, Na, and K enrichment in shales higher in the stratigraphy, suggest that mineralizing fluids upwelled from the Lamotte and Bonnette in the Viburum trend area (Gregg and Shelton, 1989a). Detailed study of the concentration of these elements in the most fractured parts of shales above the Buck mine, where upward leakage was perhaps greatest, show that enrichment of Na, Mn, Fe, and Zn was followed by depletion to background levels (Panno et al., 1988). Depletion did not occur in the less fractured parts of the shale. The leaching of the fractured parts of the shales may be related to the dissolution of sulfides and hydrothermal dolomite in the ore zones;
both could be caused by meteoric inflow. As shown in Figure 3, fluid-inclusion homogenization temperatures in different mining districts in the region range from 60°C to 180°C. Fluid inclusions in baroque dolomite in the upper Bonneterre formation in the Reelfoot Rift suggest two much hotter episodes of heating, the first with mean homogenization temperatures of 235°C and the second with mean temperatures of 170°C (Tobin, 1991). Fluid inclusions in the core of the Ouachita Mountains have even higher homogenization temperatures up to 300°C (Shelton et al., 1986). A general decrease in temperature across the ore district is suggested by these data, especially if flow was directed from the Reelfoot rift, and the Ouachita homogenization temperatures are indicative of the temperatures of fluids entering the Cambrian aquifers at the base of this sediment pile. The absence of baroque dolomite in the mining districts in Missouri and Arkansas, however, and the scatter is far bigger than the difference in modal temperature between districts. Leach and Rowan (1986) attribute this to inclusion necking, but it could also reflect real temperature variations associated with pulsing flow. The reality of the modal trend of 0.05°C/km they deduce from the mining district data in Figure 3 could certainly be questioned.

(e.g., Gregg and Shelton, 1989a; Shelton et al., 1992), and for this reason the line connecting the modal data drawn in Leach and Rowan’s original figure has been omitted. The general consensus is that the sizes and shapes of the ore deposits have been essentially preserved, and that the burial temperatures considerably above those that would be predicted by their burial depths at the time of mineralization. This is the most significant point shown by Figure 3. Almost everyone attributes a large scatter in the homogenization data from the mining districts in Missouri and Arkansas, however, and the scatter is far bigger than the difference in modal temperature between districts. Leach and Rowan (1986) attribute this to inclusion necking, but it could also reflect real temperature variations associated with pulsing flow. The reality of the modal trend of 0.05°C/km they deduce from the mining district data in Figure 3 could certainly be questioned.

THERMAL CONSTRAINTS ON FLUID FLOW

Both cross-basin hydrologic flow and basin dewaterting must produce flows sufficient to explain the fluid inclusion homogenization temperatures in Figure 3. We start with an analysis of this thermal constraint following the approach developed by Cathles and Smith (1983). We ask the simple question: How fast must waters flow out of the Arkoma Basin to heat the Cambrian aquifers in the Ozarks to temperatures of 80°C to 120°C? The critical issue is the heating of the aquifer system. If the aquifer system has the required temperature, fluids will be able to move locally upward in fractures, which cross-cutting strata and deposit minerals with little loss in temperature, recording the aquifer temperatures as they do so. If the aquifer system is not heated sufficiently, local fluid upwelling at any rate will not be able to produce fluid inclusions with the required homogenization temperatures. Put the problem in a present-day context, temperature distribution in the Arbuckle or Rice aquifers in Kansas would need to be 80°C to 120°C at 1-km depth. They are ~40°C (25°C to 30°C above ambient) at this depth today. Weak flow through these aquifers elevates the thermal gradient near where the aquifers crop out to ~40°C km⁻¹, but these gradients do not extend to depth. Much greater flow rates would be required to bring waters of 80°C to 120°C within a kilometer of the surface.

How much faster the flow must have been to accommodate the temperatures and depths thought to have existed during the Paleozoic in Missouri and Arkansas can be determined by calculating the temperature profile in a basal aquifer or
A discussion of flow mechanisms in the Cambrian aquifers of the Ouachita-Arkoma basin-Ozark system. The aquifer system by daisy-chaining analytical solutions for steady state temperatures along aquifer segments of uniform dip (Cathles, 1987). Analytical solutions for each linear uniform dip segment of the aquifer are joined together by requiring continuity of flow and temperature. The temperature at only one location along the aquifer need be specified. The temperature profile then depends on the thermal conductivity, $K$, of the sediments above the aquifer, the regional heat flow, $j_r$, and the total flow per unit strike length through the aquifer or aquifer system.

Both $j_r$ and $K$ are parameters of critical importance, especially in the flat portion of the flow path across the Ozark plateau. The Cambrian to Mississippian cover in this area consists today of a basal sandstone overlain by carbonates with some shales. In the St. Francois Mountains area, this stratigraphy is ~80% carbonate, ~12% sandstone, and ~8% shale (Figure 2).

The thermal conductivity of a lithologic sequence can be estimated using the porosity-dependent rock type thermal conductivities and porosity-depth relations in Royden and Keen (1980). I calculate using Royden and Keen’s relations that the effective thermal conductivity of a 2.5-km-thick column of sediment consisting of 80% carbonate, 12% sandstone, and 8% shale is ~5.6 × 10^3 cal/cm·s·°C. Thermal Conductivity Units (or TCU). Without compaction, the thermal conductivity would be ~5 TCU. The column would have to be 53% shale to have a thermal conductivity of 5 TCU. If the sediments now eroded from the Ozark plateaus were dominantly shale, a value as low as ~4 TCU might be approached. It is more likely, however, that these shallow water sediments would have been sands or carbonates, in which case the effective thermal conductivity of a 2.5 km section would have been between ~5.6 TCU (all carbonates) and ~6.5 TCU (90% carbonates, 10% sand). A lower bound for the thermal conductivity of the sediments covering the Ozark plateaus at the time of brine movement is therefore probably ~4 TCU and the most reasonable estimate ~5.6 TCU or greater.

Heat flow in Missouri today is ~3.3 × 10^3 cal/cm²·s or 1.3 Heat Flow Units (HFU; Kron and Stix, 1982). There is no reason for it to have been significantly different in Perman time.

Figure 5 shows the thermal profile along the “Cambrian aquifer loop” produced by cross-basin flow of various magnitudes. In Figure 5A, K=5.6 TCU, $j_r=1.3$ HFU, and $Q(\text{cal/cm}^2\cdot\text{s})$ is specified in terms of the dimensionless parameter $\alpha$, where $\alpha = K/Q$ or $\tan \beta$, and $\beta$ is the dip of the steep Ouachita-Arkoma aquifer segments (+0.1) and $c$ is the heat capacity of water ($1 \text{cal/g} \cdot \text{°C}$). Temperatures in the Cambrian aquifers (Lamotte Sandstone, Eminence Dolomite, etc.) are not elevated significantly at any flow rates for these parameter values. The initial Ouachita-Arkoma Basin complex is cooled by the high flow rates (low values of $\alpha$) before aquifers in the Ozarks can be significantly warmed. Figure 5B shows that if $K$=4 TCU, cross-basin flow can warm the Ozark Cambrian aquifers to the levels indicated by the modal fluid inclusion homogenization data in Figure 3 in the south end of the transect near the Northern Arkansas District. However, even very large uniform flow rates cannot produce the temperature increases required by the modal (let alone maximum) fluid inclusion homogenization data in the north near the Central Missouri District. Uniform cross basin hydrologic flow thus cannot match the modal fluid inclusion temperatures over the entire Ozark transect for the depth of burial selected, even in the unlikely event that the effective thermal conductivity of the cover was as low as 4 TCU. The Ouachita-Arkoma belt is too narrow to sufficiently heat waters flowing uniformly across it at the rates that would be required to warm the margin to the levels observed.

It is important to note that if the thermal conductivity of the Ozark cover could be lowered just slightly below 4 TCU, or if the cover thickened, the requisite temperatures could be attained with very slow or no fluid flow. For example, if the effective thermal conductivity of the Ozark cover were ~3 TCU and the regional heat flow 1.5 HFU, the unperturbed thermal gradient would have been 50°C/km and temperatures adequate to explain the Ozark fluid inclusion homogenization temperature data could be obtained under 1.5 km of cover with no fluid flow at all. A similar result could be obtained if the cover were substantially thicker. In either case, fluid movements could have been slow and the required Pb, Zn, Mn, etc. brought in over a protracted period of time.

If the Ozark cover were not thick and insulating, cross-basin flow could still produce the required temperatures in the Ozark aquifers if the flow were non-uniform in time or space. If the flow is sufficiently focused once it leaves the deep warm sections of the flow loop under the Ouachita Mountains and Arkoma Basin, the temperatures required by the fluid inclusion data can be attained at the Central Missouri Pb-Zn district and other distal Ozark locations. The Ouachita-Arkoma flow rates that raise the temperatures of the 3-km-deep slope break the most (to ~188°C above surface ambient) are characterized by $\alpha$~2 (see Figure 5A). If the flow is then focused by a factor of 40, daisy-chain calculations indicate that distal locations are warmed to the extent required by the modal fluid inclusion data.

The requisite distal temperatures could also be attained by transient (pulsed) flow. Across-basin flow will initially displace deep, warm brines from the Arkansas just as will compactive or other types of basin deformation. The initial flow of warm brine could heat the Ozark aquifers to temperatures.
Figure 5: (A) Temperature profile in degrees Centigrade above ambient along flow path \( a \) (which starts at the surface in the Ouachita highlands, moves to 14 km depth, then to 3 km depth and then to discharge in Ozark plateaus as shown in Figure 1C) calculated for various values of the dimensionless flow rate parameter \( a \). Smaller values of \( a \) indicate larger flow rates. Regional heat flow \( j = 1.3 \times 10^7 \) cal/cm²·sec or 3.1 HFU; the thermal conductivity of the cover, \( K = 5.6 \times 10^3 \) cal/cm²·°C or 5.6 TCU. On this and the other figures the normal unperturbed temperatures along the Cambrian aquifers are indicated by the lower heavy line labeled “unperturbed.” The modal homogenization temperature in Figure 3 with an ambient temperature of 20°C subtracted are indicated by the upper heavy line labeled “observed.” (B) Temperature profiles for various values of \( a \) along loop \( a \) in Figure 1 for \( K = 4.7 \) TCU and \( j = 3.1 \) HFU. (C) Temperature profile (in degrees Centigrade above ambient) along flowpath \( b \) (which starts at 14 km depth and then follows flowpath \( a \) as shown in Figure 1C) calculated for various values of \( a \) under the same conditions as Figure 5A. Fluids are assumed to enter the Cambrian aquifers at 14 km depth at equilibrium temperatures (depth 32°C above ambient). The flow rates, \( Q \), shown in this figure correspond to the values of \( a \) given in Figures 5A and 5B. For example, \( a = 0.1 \) corresponds to \( Q = 1744 \) km³/km²·y. The formula relating \( a \) and \( Q \) is given in the text.

approaching those in the deepest parts of the Ouachita-Arkoma basin. With continued flow, those deep locations will be cooled by cold meteoric inflow and temperatures in the Ozark aquifers will drop. Temperatures in the Ozark aquifers will thus rise and fall in a transient fashion. The maximum transient temperatures that could be attained at various flow rates can be estimated if brine is introduced to the deep extension of the Ozark Cambrian aquifers at temperatures normal at those depths under no flow, thermal equilibrium conditions (path b in Figure 1B). Figure 5C shows that such flow can elevate temperatures in the Cambrian aquifers to the highest end of the modal ranges indicated in Figure 3. The required temperatures in the south can be produced by flow rates of 24 km³/km²·y. In the north, flow rates of 1764 km³/km²·y. (\( a = 0.1 \)) are required. Temperatures in the south and north are thus compatible with the flow in the Cambrian aquifers that concentrates by a factor of 4 as it proceeds north. If the aquifers in question have a cumulative thickness of 100 m, the Darcy flow rate would have increased northward from 4.4 to 17.6 m²/s.

There is general agreement regarding the flow rate required to heat a basin margin. For example the flow rates required to significantly warm the outflow margin of the Illinois Basin (slope \(-0.75\%\)) in Bethke’s (1986, Figure 1b) analysis of the Upper Mississippi Valley Pb-Zn district in Wisconsin was \(-540 \) km³/km²·y. The long, flat, deep flow segment across the Illinois Basin avoids the heating problems caused by the narrowness of the Ouachita-Arkoma system and allows sufficiently warm fluids to enter the basin. It is only when temperatures are too high that warming enough fluids are obtained, however, the flow rates needed to warm the margin aquifers are similar to those we estimate. The fundamental parameter controlling margin heating is \( Q \). Our estimates of \( Q \) are minimum estimates. To the extent that flow is less concentrated, the possibility of thermal losses being greater is uncertain. In that case \( Q \) would be greater, because the heat has less distance to diffuse to the surface, and \( Q \) must be greater for the same basin margin temperature perturbation.

**SALINITY CONSTRAINTS ON CUMULATIVE FLUID FLOW**

Requisite flow rates of \(-441 \) km³/km²·y. and the present-day pore water salinity distribution place severe constraints on how possible flow mechanisms could have operated in the Ouachita-Arkoma-Ozark area. In cross-basin hydrologic flow, fresh meteoric water will rapidly displace brines and dissolve any salt contacted. For example, if the Arkoma-Ouachita sedimentary basin (10 volume% brine mixtures and the flow contacted these evaporites, cross-basin flow at \(-141 \) km³/km²·y. would dissolve and carry away all the salt in 1.25 m.y. In this calculation we assume 10% of the 720 km³/km² of Ouachita-Arkoma section is salt, that salt has a density of \( 2 \) g cm⁻³, and that the efflux has a salt-saturated salinity of \( 26 \) wt%. The salt is dissolved (and the temperature of the basin in time \( t \) is \( 0.26 \) Q, equals the amount of salt in the basin, 720 (0.1) (2 cm³⁻¹). If the flow rate were \(-88 \) km³/km²·y. (\( a = 2 \), the same amount of salt would be flushed in 0.9 m.y.

The dilemma such fast flushing of salinity poses for cross-basin hydrologic flow is as follows. Erosion is a slow process. Once the topography to “drive” cross-basin hydrologic flow is established, it will take tens of millions of years to remove. The permeability of the MVT deposits we have been discussing remains very high even after years. Today, six and a half tons of water were pumped for every ton of ore mined in the Old Lead Belt in 1943 (Weigel, 1945). Two years of pumping were required to achieve a 700 ft drawdown in the Viburnum Trend (Bullock, 1973). Thus, although there may have been permeability reduction where metals were deposited in disseminated mineralization in matrix pores (\(-90\% \) of the ore is of this type; personal communication from Jay Gregg, 1992), the fracture and breccia pathways that channeled fluids to these areas remains high. The flow of fluids required to produce low brine density through groundwater evaporation or snowmelt is large. Also it is clear from the position of the deposits that brine discharged upward through them to the surface at the time of mineralization. Consequently, the amount of fluid required to flow brines could remain in the basins surrounding the Ouachitas if cross basin hydrologic flow at the rates required to warm the margin aquifers at these flow rates and brine salinities in MVT deposits in the area were ever established.

The lack of a full (or even partial) salinity flush from the deep parts of the basins surrounding the Ozark plateau that are thought to have produced the MVT mineralization in the region (the Illinois and Arkoma basin) appears to pose a fundamental difficulty for the cross-basin hydrologic flow model in the Ouachita-Arkoma-Ozark area and elsewhere (Cathles, 1987). Either a way must be found to turn off cross-basin hydrologic flow shortly after it is established by tec-tonic events, or a way must be found to explain why the basin brines have not been flushed by such flow.

Compactive dewatering does not have a salinity flush problem because fresh water does not enter the basin and only a portion of the pore waters are compactively expelled. On the other hand, as discussed by Cathles and Smith (1983), transient compactive expulsion is constrained by two kinds of heat balance considerations. If the accumulation of the 720 km³/km² of sediments shown in Figure 1C occurred in 10 m.y. (probably a generously rapid rate), for example, and caused 10% compaction, fluids would be expelled at \(-7 \) km³/km²·y. This is \(-1.6\% \) of the rate required to warm the margins (441 km³/km²·y.) by the requisite amount. Dewatering of the Arkoma must be focused \(-62 \) fold on the margins or temporally episodic if the dewatering model is to warm the Ozark Plateau Aquifers as observed. Furthermore, it is proposed that compactive dewatering pulse must be of sufficient volume to warm the escape aquifers. If the Cambrian Ozark aquifers cumulatively had an average thickness of 100 m over the 330 km from their bend in Figure 1B to the Central Missouri MVT District, a volume of water \(-33 \) km³/km² would be required for each pulse. The 720 km³/km² sediment prism in Figure 1b could thus contribute about 2 pulses of fluid. If there were progressive channeling of brine flow paths to the north, which is reasonable to expect and indicated by the pattern of geochimical alteration (channeling along the El Dorado Reef foot), more pulses could be delivered. For example, if flow were focused by a factor of four by channeling, 8 pulses of warm brine could be delivered to areas as far north as Central Missouri.

**FLOW REQUIREMENTS OF MINERALIZATION AND ALTERATION**

The limited quantities of brine that might be compactively expelled from the Ouachita-Arkoma system are too small to account for the observation of Arkansas and Missouri. If we take the axis of the Broken Bow uplift as the southern boundary of the sediment volume that might contribute fluids to the Ozark Plateau (with the logic that any sediments south of this divide would expel their fluids south, not north), then the Ouachita-Arkoma sediment volume is approximated by a triangular wedge 12 km deep and 120 km wide (Figure 1C). The volume of such a wedge is 720
The calculations indicate the importance of parameters perhaps not emphasized enough in the basin modeling literature. The thermal conductivity of margin cover is critical because of the Old Yellow Belt on the eastern side of the St. Francois Mountains produced about 9 x 10^10 tonnes of lead (Snyder and Gerderman, 1968). The Tri-State District produced about 14 x 10^10 tonnes of zinc (20%) (Brodie and others, 1968). The other Central Missouri and Northern Arkansas mining districts contained smaller amounts of metal. All told, the mineralization in Arkansas, Missouri, and immediately adjacent areas totals perhaps 8 x 10^10 tonnes of combined Pb and Zn.

Bassin brines can easily contain ~20 ppm dissolved Pb or Zn (Carpenter and others, 1974). If just 10 ppm lead plus zinc precipitated from basin brines propelled across the Ozarks, 8000 km^2 of brine would be required to precipitate the known resources of the area. This volume of brine represents a brine-filled porosity of less than 1.5% of the 566,000 km^2 of sediments defined above. The sediment volumes identified would have had little difficulty supplying the brine-generated mineralization. Quantitative discussion of how basin brines behave as ore fluids has been offered by Sverjensky (1984).

The brines identified sediment volumes are also sufficient to produce the alteration observed in the Cambrian aquifers of the Ozarks. Gregg (1989) pointed out the extensive nature of the epigenetic dolomitic cements at the base of the Bonnereter. He estimated that ~100 km^2 of Bonnetter limestone was dolomitized over an area of ~16,600 km^2 area near the St. Francois Mountains and that 35,000 km^2 of brine was generated by this dolomitization. Even this seemingly very large brine volume represents the brine in pores representing just 6% of the 566,000 km^2 sediment volume. Clearly either cross-basin topographically driven flow or compaction could relatively easily supply the volumes of brine required for alteration and mineralization without exhausting the supply of brine in the Ouachita-Arkoma system.

DISCUSSION, SUMMARY, AND CONCLUSIONS

The calculations presented above are selected to make several specific points. Many different combinations of flow path, focusing, overburden thickness, thermal conductivity, and heat flow could have been selected. The calculations are simple enough using the stream of information on hand that a reader can easily explore these possibilities and, I hope, will be encouraged to do so by this discussion.

Alteration and mineralization. One argument against episodic brine expulsion is that the Mid-Continent deposits do not appear to have occurred many times at the time of ore deposition. The calculations presented in Figure 5 also clearly show that the deposits need not have been local temperature anomalies. Whether or not this is so will depend on localized escape from the Cambrian aquifers and per (near the paleo-surface) the deposits were in the stratigraphic column at that time. If the deposits were near the paleo-surface under ~1 km of cover, local thermal anomalies would be lost in the scatter of the homogenization data, and the most obvious sign would be the elevation of temperatures with perhaps some hint of a slight up dip decrease in the mode of the homogenization data.

Compactive dewatering might have been assisted by a rise in brine density in the Arkoma basin almost entirely filled with gas) raises the intriguing possibility that gas generation may have expelled brine, a possibility suggested long ago in general and specifically for the Ouachitas by Rich (1927). Rates of gas-driven brine expulsion are competitive with compaction for moderate kergen generation times. This makes good sense if I am right that the gas generation or gas-driven fluid expulsion occurred in pulses, as particularly organic rich strata were heated or as seals ruptured, for the rate of expulsion of gas can be much higher than for the bulk basin hydrologic flow. Regardless of their cause, pulses of fluid flow are suggested by a variety of observations and have geochemical advantages when it comes to alteration. For example, Sverjensky (1981) documents at least eight major pulses of mineralization interspersed with periods of ore dissolution in the Buick mine of the Viburnum trend. At least five events of chlorite, pyrite, and garnet deposition, three intervals of sphalerite deposition, and five intervals of galena and quartz deposition in the Tri-State district of Kansas, Oklahoma, and Missouri (p. 487).

The deposition of metallic sulfides took place over a long period of time. This condition is attested to by the many repetitions in the paragenetic sequence... and most isotopic signatures of Pb isotopes recorded from the center to the edge of a single crystal... The sulfides were even subject to periods of corrosion between periods of deposition... The corrosion could be produced by meteoric recharging of the Cambrian aquifers as is happening today between pulses of brine expulsion. The low Fe and Mn contents in the back reef areas of the Viburnum Trend suggest this recharge (Buelter and Guillemette, 1988). The pressure-temperature-incompatible mineralization (Barton, 1981) and stable isotopes (Richardson and others, 1981) also indicate pulses of fluid flow with minerals deposited at different temperatures along perhaps many prograde-retrograde thermal gradients. Pulses of fluid flow can achieve a cumulatively greater alteration because alteration occurs many times at the time of ore deposition. The calculations presented in Figure 5 also clearly show that the deposits need not have been local temperature anomalies. Whether or not this is so will depend on localized escape from the Cambrian aquifers and per (near the paleo-surface) the deposits were in the stratigraphic column at that time. If the deposits were near the paleo-surface under ~1 km of cover, local thermal anomalies would be lost in the scatter of the homogenization data, and the most obvious sign would be the elevation of temperatures with perhaps some hint of a slight up dip decrease in the mode of the homogenization data.

Some aspects of the fluid inclusion data just briefly mentioned in the literature may provide critical evidence. Gregg and Shelton (1989a) point out that early fluid-inclusion studies by Rooder and later studies by Bauer found a high-temperature, low-salinity population of fluid inclusions. Fluid inclusions at any given homogenization temperature in fact range from low to high (brine) to very low salinity inclusions occur in the dolomite cements from late diagenic through ore deposition time (Shelton et al., 1992).

There are at least three possible explanations for the low salinity inclusions. They could record pulses of metamorphic fluids from the core of the Ouachita Mountains. Fluid inclusions with homogenization temperatures up to 300°C are common. NaCl equivalent are found in Ba-rich adularia in organic-rich seams in Cambrian shales on the shores of Lake Ouachita in Arkansas (Shelton et al., 1988). Alternatively they could record pulses of basinal hydrologic flow. In either of these two cases, pulses of low salinity fluid would flush brine out of flow channels and trap low salinity fluid inclusions. Brines would re-enter the flow channels from less permeable zones after the low salinity and heat pulses had passed. Finally the lower salinity inclusions could be produced by the flushing of lower salinity fluids recharged by pre-existing lower salinity fluids.

This in fact seems to be the most likely explanation. The 25,000 km^2 study area in southeastern Missouri that Gregg and others (1989) analyzed, and which now provides the best documentation of the lower salinity inclusions, lies entirely within those parts of the Ozark Plateau where the rock has been displaced by brine (dashed box in Figure 4). Thus most if not all of the samples containing lower salinity inclusions could have been affected by meteoric recharge between pulses of brine expulsion. The isotopic composition of the low salinity fluid inclusions and their abundance patterns should distinguish these possibilities. Other models presented above and additional testable explanations for the lower salinity fluid inclusions indicate the importance of fluid focusing. Geologic observations suggest some focusing of flow in the Reelfoot Rift. Groundwater collected above steady cross-basin topographically-driven flow and the modal fluid inclu-
A discussion of flow mechanisms in the Cambrian aquifers of the Ouachita-Arkoma basin-Ozark system


