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Hydrologic Constraints on the Genesis of the Upper Mississippi Valley Mineral District from Illinois Basin Brines

CRAIG M. BETHKE

Hydrogeology Program, Department of Geology, 245 Natural History Building, University of Illinois, Urbana, Illinois 61801

Abstract

Mississippi Valley-type deposits of the Upper Mississippi Valley mineral district probably formed during a period of regional ground-water flow across the Illinois basin initiated by uplift of the Pascola arch in post-Early Permian and pre-Late Cretaceous time. Numerical modeling of this inferred paleohydrologic regime shows that district temperatures attained by this process depend on flow rates through the basin, heat flow along flow paths, and presence of structures to cause convergence and upwelling of fluids. Predicted flow rates and timing of mineralization agree with previous estimates. Modeling results also offer explanations of banding in district mineralization and district silicification patterns. Modeling of ground-water flow due to sediment compaction during basin subsidence, however, shows that this process was not responsible for mineralization. Fluids displaced from the deep basin by compaction-driven flow moved too slowly to avoid conductive cooling to the surface before reaching the district. Episodic dewatering events are unlikely to have occurred, because the basin did not develop significant overpressures during subsidence. Results of the compaction-driven flow modeling probably also apply to the Michigan and Forest City basins. Study results suggest that exploration strategies for Mississippi Valley-type deposits should account for tectonic histories of basin margins distant from targets.

Introduction

ORES of the Upper Mississippi Valley, or Wisconsin-Illinois, mineral district are well-studied examples of Mississippi Valley-type ore deposits. Mississippi Valley-type deposits contain epigenetic, hydrothermal ores, both as open-space-filling and replacement mineralization. The ores are commonly found in shallowly buried rocks on margins of sedimentary basins (Anderson and Macqueen, 1982). Fluid inclusions from these deposits typically give temperature estimates of 100° to 150°C and are filled with predominantly Na-Ca-Cl brines containing >15 wt percent salts (Roedder, 1979a, p. 95). Mississippi Valley-type ore-forming fluids differ significantly in chemical and isotopic composition from fluids which form other types of hydrothermal deposits (Roedder, 1979b), but many workers have compared these fluids to modern oil-field brines (White, 1958; Hall and Friedman, 1963; Giordano, 1978; Taylor, 1979; Sverjensky, 1984). This similarity is partly responsible (Ohle, 1980) for the present popularity of the basinal brine theory of Mississippi Valley-type genesis, suggested by White (1958). According to this theory, warm sedimentary brines from nearby basins carry dissolved metals and possibly sulfur from depth to basin margins, forming ore deposits.

Three variants of the basinal brine theory currently exist. The "stratifugic" fluid variant holds that deep fluids move toward basin margins as a result of sediment compaction during basin evolution (Noble, 1963; Jackson and Beales, 1967; Dozy, 1970). In a second, episodic dewatering variant, sediment compaction in basins causes high excess pore pressures, such as those observed in the U. S. Gulf Coast (Dickinson, 1953), which drive sudden bursts of deep brines toward basin margins (Sharp, 1978; Cathles and Smith, 1983). A third variant, suggested nearly a century ago (Daubree, 1887; Cox, 1911; Siebenthal, 1915) and recently revived and revised by Garven and Freeze (1984a and b) and Garven (1985), holds that gravity-driven ground-water flow due to topographic differences across basins carries warm fluids from deep strata into shallow sediments. Evaluating each variant is important to understanding relationships of parent basins to mineralization and refining exploration strategies for Mississippi Valley-type deposits.

This paper uses numerical modeling techniques to study possible hydrologic mechanisms of forming the Upper Mississippi Valley district from brines derived in the Illinois basin. The Illinois basin is the most plausible source of ore-forming fluids because of its proximity to the district, its depth, and the presence

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of mineralization within the basin which seems genetically related to the Upper Mississippi Valley district. Modeling of compaction-driven ground-water flow shows that the stratifugic fluid variant is not viable because fluids move slowly and cool by conduction long before reaching host rocks at the basin margin. Calculations also show that, due to slow burial rates and low shale content, the Illinois basin was not overpressured during its compaction. This result indicates that episodic dewatering events are unlikely to have occurred. Other possible fluid sources, such as the Michigan and Forest City basins, are tectonically similar to the Illinois basin and also seem unlikely to have formed the Upper Mississippi Valley district by compaction flow.

Models of gravity-driven ground-water flow, however, demonstrate that this process, possibly resulting from tectonic events in the southern basin, is capable of having introduced warm fluids to the Upper Mississippi Valley district. Gravity-driven flow due to topographic differences across basins is the preferred variant of the basinal brine theory in the Illinois basin, and by inference in other intracratonic environments. This result has important implications for Mississippi Valley-type exploration.

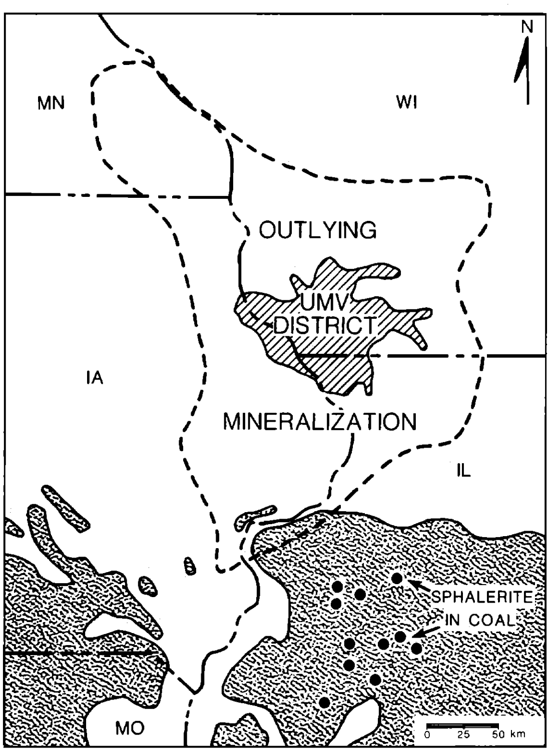

FIG. 1. Location of Upper Mississippi Valley mineral district, showing mining district, area of outlying mineralization, and occurrences of hydrothermal sphalerite in Pennsylvanian coals (after Cobb, 1981; Heyl and West, 1982). Shaded areas show locations of Pennsylvanian rocks of the Illinois and Forest City basins.

Upper Mississippi Valley Mineral District

The Upper Mississippi Valley mineral district (Fig. 1) is a widespread ore district covering 8,000 km^2 in a thin (<1 km thick) veneer of Cambrian through Silurian strata overlying the Wisconsin arch in Wisconsin and Illinois (Heyl et al., 1959; Heyl, 1968). The district contains primarily zinc and lead ores, mostly deposited as open-space fillings, in pitches and flats, tectonic and solution breccias, veins, joints, and vugs. Heyl et al. (1959) identified silicified and mineralized veins and fractures extending into upper Cambrian sandstones, the stratigraphically lowest sediments in the Illinois basin, as the "mineralized roots" of the ore district.

Noneconomic mineralization occurs over a much larger, 100,000 km^2 area in Wisconsin, Illinois, Iowa, and Minnesota. The crystal habit, paragenesis, and trace element and isotopic composition of much of this mineralization is similar to Upper Mississippi Valley ores (Heyl and West, 1982; Garvin et al., 1985). Epigenetic sphalerite also occurs in Pennsylvanian coals and Mississippian-Pennsylvanian concretions on the western shelf of the Illinois basin. Cobb (1981, p. 181–186) interprets these occurrences to be genetically related to district ores.

Although located in shallow sediments, the district was extensively heated during mineralization. Fluid inclusions from ore minerals give estimates of temperatures during deposition of 75° to 220°C (Newhouse, 1933; Bailey and Cameron, 1951; McLimans, 1977). These values agree with temperatures of 52° to 227°C calculated from sulfur isotope compositions of coprecipitated galena and sphalerite (McLimans, 1977, p. 109–131). Measurements of fluid inclusions from sphalerite in coal give estimates of 75° to 113°C, and vitrinite reflectance of host coals also suggests heating to 70° to 110°C (Cobb, 1981). Fluid inclusions in diagenetic cements from shallow (<450 m deep) Mississippian strata in western Illinois give estimates of 85° to 109°C (Smith et al., 1984).

McLimans et al. (1980) show that color banding in district sphalerites forms a distinct paragenetic stratigraphy which can be traced for many kilometers in some cases. They argue that paragenetic uniformity and textures of ore minerals are consistent with a slow, district-wide mineralizing process in which metals and sulfur were transported by the same fluid. Assuming that the zinc distribution in carbonate host rocks is due to diffusion from ore-forming fluids, Lavery and Barnes (1971) estimate that mineralization occurred over a period of about 0.2 to 0.25 m.y. Gize et al. (1981) note that the thermal maturity of organic matter associated with Upper Mississippi Valley mineralization is consistent with these interpretations of the degree and duration of heating in the district.

The age of mineralization in the district is poorly constrained (Heyl, 1968, p. 442), but the most probable interval for ore emplacement is Pennsylvanian to Cretaceous (McLimans, 1977, p. 24; A. Heyl, pers. commun.). This interval agrees with oxygen isotope and fission-track dating studies of basement rocks in the southern district (Zimmerman, 1981; Shieh, 1983) which suggest a period of hydrothermal alteration ending 130 ± 14 m.y. ago (Early Cretaceous). Mineralization postdates lithification of Silurian rocks and probably occurred after an inferred period of regional deformation during the Pennsylvanian. Sphalerite in Pennsylvanian coals was emplaced after the coals were cleated (Cobb, 1981, p. 125–130). McLimans (1977, p. 73) also points out that adequate sediment thickness to prevent ore fluids from boiling, assuming deposition in shallow water, was not present until Mississippian time. Richards et al. (1972), however, reinterpreted J-type lead isotope data from the district to be consistent with Middle Devonian to Late Pennsylvanian mineralization.

Illinois Basin Tectonic History

The Illinois basin is an intracratonic basin (Sleep et al., 1980) containing gently dipping Paleozoic strata in Illinois, Indiana, and Kentucky (Fig. 2). Perhaps to a greater degree than other intracratonic basins in North America, such as the Michigan basin, the Illinois basin has had a complex tectonic history. During most of the Paleozoic, the basin was apparently trough shaped and trended north-northwest to south-southeast. Subsidence and sedimentation began in Late Cambrian time and continued intermittently at least into the Pennsylvanian. Six major unconformities occur within this interval (Willman et al., 1975). In post-Early Permian time, uplift of the Pascola arch and later subsidence of the Mississippi embayment cut off the southernmost one-third of the basin. Although the southernmost basin is generally assumed to have mirrored its preserved counterpart to the north, there is little direct evidence of its past structure, and other interpretations have been proposed (Swann, 1967).

At the deepest point, more than 4 km of Paleozoic strata overlie a Precambrian crystalline basement, which is probably more than 1.1 b.y. old (Grogran, 1950; Bradbury and Atherton, 1965; Lidiak et al., 1966). Cambrian and Early Ordovician sediments are dominated by mature sandstones such as the Mt. Simon (<1 km thick) and St. Peter (<250 m) quartz arenites. Because of their stratigraphic position and hydraulic continuity (Bond, 1972), these sandstones are often referred to as basal aquifers. Overlying Late Ordovician through Mississippian sediments form a thick (<1.7 km) sequence of carbonate rocks, interrupted by the Ordovician Maquoketa (<110 m) and Devonian-Mississippian New Albany (<125 m) shale groups. These strata are covered by less than 1 km of Pennsylvanian cyclothems of sandstone, shale, limestone, and coal (Willman et al., 1975). The youngest Paleozoic sediments in the northern basin are of Pennsylvanian age, but Damberger (1971) suggested that 1.5 km of Permian sediments in southern Illinois would explain coal rank at present geothermal gradients, and some Permian sediments have been identified in grabens in Kentucky (Schwalb, 1982).

Geologic evidence in the southern basin suggests a postdepositional interval of high heat flow. Mafic dikes and sills in southern Illinois give Permian isotopic dates (Zartman et al., 1967) and monazite from the Hicks Dome cryptovolcanic structure described by Trace (1960) has been dated as late Cretaceous (McGinnis et al., 1976). Mesozoic plutons have been inferred from gravity and magnetic studies on the western flank of the Pascola arch (Schwalb, 1982, p. 40) and in southern Illinois (McGinnis et al., 1976). Past high heat flow would also explain anomalous coal ranks in this area (Damberger, 1971).

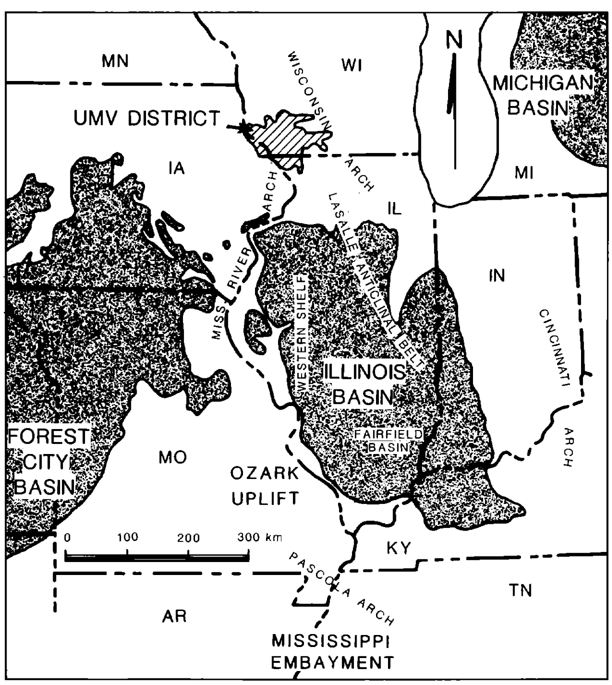

FIG. 2. Structural setting of Illinois basin and Upper Mississippi Valley district (after Heyl et al., 1959; Bond et al., 1971). Areas of outcrop of Pennsylvanian rocks are shaded.

Paleohydrologic Models

This paper presents two paleohydrologic models of the Illinois basin. The first considers compaction-driven ground-water flow during basin subsidence beginning in Late Cambrian time and continuing through deposition of hypothesized Permian sediments. The second model simulates gravity-driven ground-water flow due to subsequent uplift of the Pascola arch during tectonic deformation of the southernmost basin.

Compaction-driven ground-water flow

In order to investigate relationships of compaction-driven ground-water flow during subsidence of the

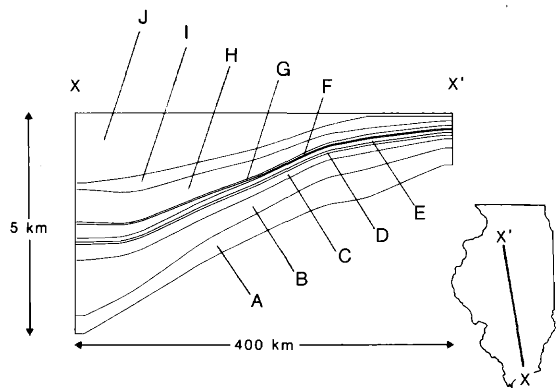

FIG. 3. Basin cross section used for compaction-driven flow calculations. Vertical exaggeration is 45:1. Cross section extends 400 km from Fairfield basin north-northwest toward the Upper Mississippi Valley district. Stratigraphic units A-J are summarized in Tables 1 and 2.

Illinois basin to genesis of the Upper Mississippi Valley mineral district, a compaction-driven flow simulation was made using numerical modeling techniques described by Bethke (1985b) and summarized in the Appendix. The mathematical model considers fluid flow due to pore volume collapse and thermal expansion of pore fluids during burial, as well as heat transfer by conduction and by advecting ground waters. The calculation was made along a cross section extending from the Fairfield basin, the present thickest part of the Illinois basin, north-northwest for 400 km to the west of the LaSalle anticlinal belt and east of the Mississippi River arch (Fig. 3). This cross section was split into ten time stratigraphic divisions based as closely as possible on basin rock stratigraphy (Table 1). The uppermost stratigraphic division, as much as 1.5 km of Permian sediments postulated by Damberger (1971), was included to achieve maximum effect from compaction-driven flow.

Stratigraphic divisions were taken to be interlayerings of sandstone, shale, and carbonate rocks (Table 2). Sediments were assumed to compact with depth (Fig. 4) according to results of previous studies of sandstones and shales (Perrier and Quiblier, 1974) and carbonates (Halley and Schmoker, 1983). Sandstones were assumed to be more permeable than carbonates, which were more permeable than shales, and all sediments were assumed to lose permeability with compaction (Fig. 5). Assumed permeabilities compare well with available data from Illinois basin sandstones (Buschbach and Bond, 1974; Becker et al., 1978) and from shales in subsiding basins (Neglia, 1979, p. 582).

Calculations used fluid properties of a 0.5 molal NaCl solution, approximately seawater salinity, because processes by which pore fluids are concentrated are poorly understood and not amenable to modeling. This assumption is not believed to affect results of compaction-flow modeling significantly, and possible effects on calculations of gravity-driven flow are noted below. Pore fluid densities, coefficients of compressibility and thermal expansion, viscosities, enthalpies, and heat capacities were taken from equations of state and tabulations in Phillips et al. (1980, 1981). Thermal conductivities were estimated as a function of porosity from data of Sclater and Christie (1980). Rock en-

TABLE 1. Time Stratigraphic Divisions

Division	Time stratigraphy: System	Time stratigraphy: Series	Rock stratigraphy
J	Permian		Postulated Permian sediments
I	Pennsylvanian		All Pennsylvanian formations
H	Mississippian	Valmeyeran and Chesterian	Pope Mg, Remaining Mammoth Cave Mg, Top of Knobs Mg to north
G	Mississippian Devonian	Kinderhookian Late Devonian	Basal Mammoth Cave Mg to south, Kinderhookian Knobs Mg Upper Devonian New Albany Shale Gp (Knobs Mg)
F	Devonian	Early Devonian and Middle Devonian	Basal New Albany Shale Gp (Knobs Mg) to north, Devonian Hunton Mg
E	Silurian	Alexandrian and Niagaran	Silurian Hunton Mg
D	Ordovician	Cincinnatian	Ordovician Maquoketa Shale Gp, remainder of Ottawa Mg to south
C	Ordovician	Champlainian	Ottawa Mg, Glenwood Fm, St. Peter Ss
B	Ordovician Cambrian	Canadian Croixan	Prairie du Chien Gp Jordan Ss, Eminence Fm, Potosi Dol, Franconia Fm
A	Cambrian	Croixan	Ironton Ss, Galesville Ss, Eau Claire Fm, Mt. Simon Ss

Abbreviations: Dol = Dolomite, Fm = Formation, Gp = Group, Mg = Megagroup, Ss = Sandstone

TABLE 2. Assumed Compositions of Stratigraphic Divisions

Division	Sandstone	Shale	Carbonate
J	40	60	
I	40	60	
H	10–20	10–30	50–80
G	10	80	10
F	20–30		70–80
E			100
D	10	80	10
C	20–30	10–20	50–70
B	20–40	20	40–60
A	70–80		20–30

In compacted volume percent

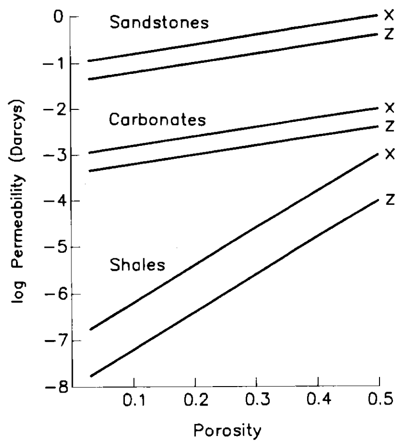

FIG. 5. Assumed permeabilities along (x) and across bedding (z) for sandstones, shales, and carbonates, versus degree of compaction.

thalpies and heat capacities were calculated with Meyer-Kelley equations for individual minerals (Robie et al., 1979) in idealized sandstone, limestone, and shale. Bethke (1985a) gives further details of assumed fluid and rock properties.

The basement contact in the simulation was defined as an impermeable barrier, and the south end of the cross section was chosen as a no-flow symmetry plane. Both the top surface and north boundaries were left open to flow. A normal continental heat flux (Lee and Uyeda, 1965) of 1.5 HFU (63 mW/m^2) was supplied across the lower bound, and the upper surface was held at 20°C. As an initial condition, the lowermost stratigraphic unit was set to hydrostatic pressure and a conductive temperature profile.

Figure 6 shows calculation results at three time levels. Arrows within domains show fluid true (or average microscopic) velocities relative to the subsiding sediments; vertical arrows along the lower bound represent subsidence velocities of the medium. Velocity vectors are exaggerated vertically to the same degree as the cross section in order to reflect flow directions accurately. Fine lines separate stratigraphic units. Bold lines are equipotentials (atm), which contour excess pressures resulting from compaction. Flow directions are not normal to equipotentials, because the permeability of the interlayered medium is anisotropic.

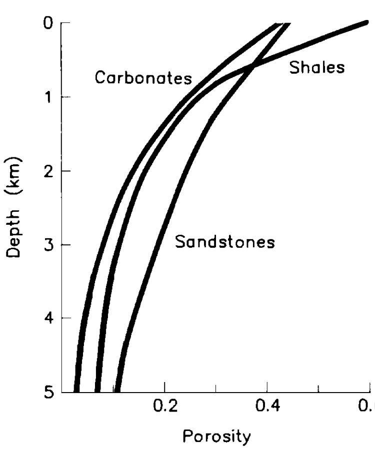

FIG. 4. Assumed porosity versus depth for sandstones, shales, and carbonate rocks, from Perrier and Quiblier (1974) and Halley and Schmoker (1983).

Compaction-driven fluids move at less than 2.3 mm/yr throughout the simulation, although these flow rates represent significant volumes over the geologic time periods considered. Small flow rates result from the very low mean burial rate of 0.002 cm/yr for the Illinois basin (McGinnis et al., 1976), which is similar to burial rates of 0.001 to 0.003 cm/yr for intracratonic basins in general (Nalivkin, 1976; Schwab, 1976). Previous study (Bethke, 1985b) has shown that, unlike gravity-driven ground-water flow in which flow rate is proportional to permeability, compaction-driven flow velocities in basins which do not develop significant excess pressures may not be increased by increasing permeability. This result arises because fluid velocities represent rates of pore volume collapse which, except in overpressured basins, do not depend on permeability. For example, the amount of fluid displaced while compacting a sediment would not be increased if the sediment were more permeable.

Figure 7 shows the temperature distribution resulting from the compaction-driven flow calculation. Isotherms, shown as solid lines, are horizontal, indicating that conductive heat flow from the lower crust to the surface has not been perturbed by fluid movement. This result is due to very small flow rates which allow fluids to cool by conduction as they move into shallower rocks. Cathles and Smith (1983) reported a similar result.

In order to test the possible effects of intermittent sedimentation at higher burial rates, the simulation was repeated with basin evolution occurring in one-tenth the actual time interval. This calculation also

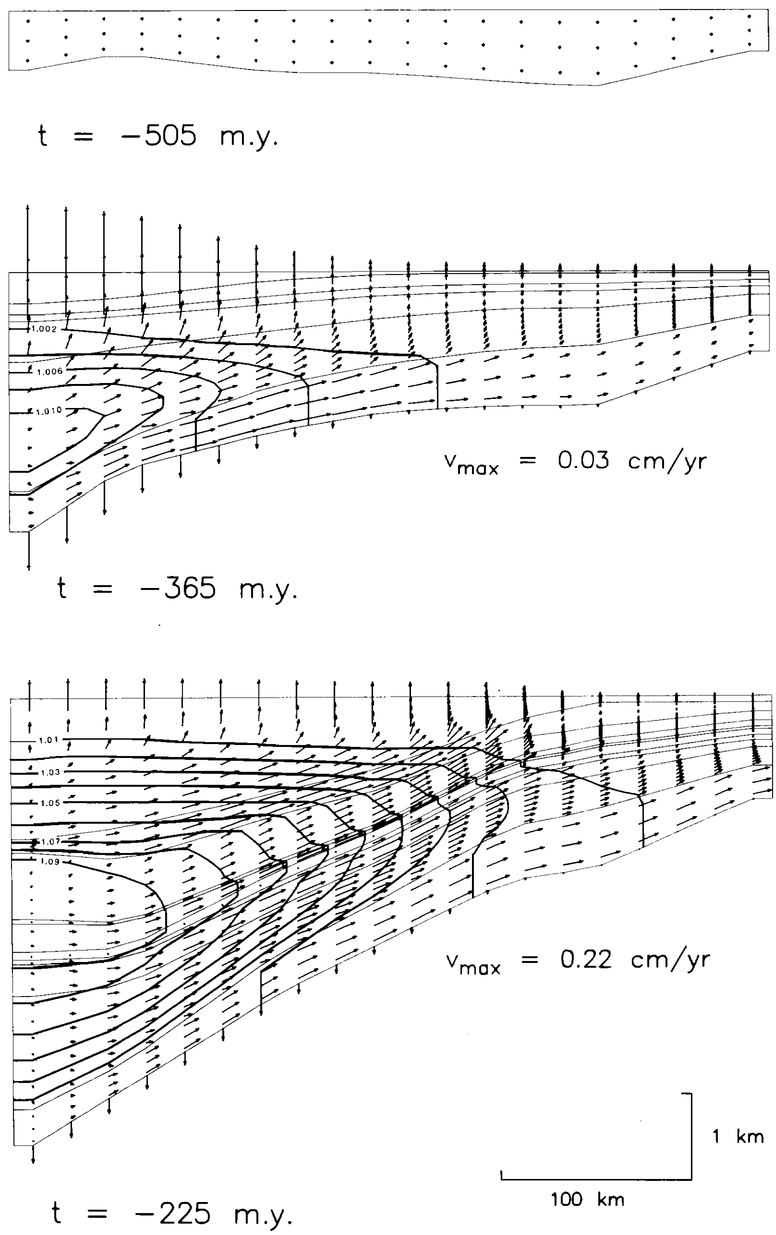

FIG. 6. Results of compaction-driven flow calculation at three time levels. Arrows show fluid true velocities relative to the medium. Vertical arrows at bottom give medium subsidence velocities. Bold lines are equipotentials (atm.); fine lines separate stratigraphic units. Intervals between equipotentials are 0.002 atm (0.0002 MPa, t = −365 m.y.), and 0.01 atm (0.001 MPa, t = −225 m.y.).

showed no tendency for advecting fluids to redistribute heat within the basin. Failure of compaction-driven ground-water flow to carry warm fluids from the deep basin toward the margin argues against genesis of the Upper Mississippi Valley district by the stratifugic fluid hypothesis.

A further result is that maximum excess hydraulic potentials of only 0.1 atm (0.01 MPa) were achieved during subsidence. For comparison, the potential gradient in a lithostatically pressured sediment is about 130 atm/km (13 MPa/km). This marked contrast with overpressured basins such as the U. S. Gulf Coast is due to both low shale content and small burial rates in the Illinois basin. While, by conservative estimate, Gulf Coast-type basins are 85 percent shale (Boles and Franks, 1979), the Illinois basin is poor in shale, especially in pre-Pennsylvanian rocks. In addition, basal aquifers in the Illinois basin provide extensive,

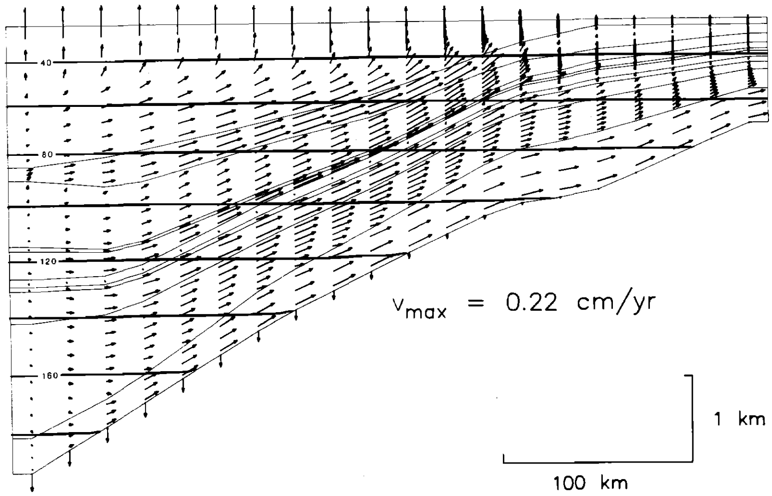

FIG. 7. Temperature distribution resulting from compaction-driven flow calculation at the final time level (t = −225 m.y.) in Figure 6. Bold lines are isotherms in 20°C increments. Fluids do not move rapidly enough to disturb the normal conductive temperature profile.

laterally continuous escape paths for fluids which are not common in the Gulf Coast. Typical burial rates in Gulf Coast-type settings of 0.01 to 1.0 cm/yr (Sharp and Domenico, 1976) are as much as three orders of magnitude greater than burial rates in intracratonic basins.

Lack of significant excess hydraulic potentials during evolution of the Illinois basin suggests that the Upper Mississippi Valley district is unlikely to have formed as a result of an episodic dewatering mechanism. Such a mechanism, however, may be important in rapidly subsiding, shaly basins such as the Ouachita basin studied by Sharp (1978) or the Gulf Coast.

Gravity-driven ground-water flow

Uplift of the Pascola arch in post-Early Permian time and persisting until subsidence of the Mississippi embayment in Late Cretaceous time (Schwalb, 1982, p. 40–43; Marcher and Stearns, 1962; Stearns and Marcher, 1962) probably caused an important gravity-driven ground-water flow regime in the Illinois basin. This structure, today only represented in subcrop below the embayment, was a large, roughly domal uplift (Fig. 8) located between the present-day Nashville dome and the Ozark uplift. The arch affected more than 40,000 km^2 and was comparable in size to the present Nashville dome, Black Hills, and Llano uplift.

Marcher and Stearns (1962) estimated minimum past elevation of the arch to be 200 m near its structural center, based on studies of sediment transport into the Cretaceous Tuscaloosa Formation in west-central Tennessee (Fig. 9). Approximately 2.5 km of sediments were eroded from the arch, exposing sediments as old as Cambrian, including Illinois basin basal aquifers such as the St. Peter and LaMotte (facies equivalent of Mt. Simon) Sandstones. Past elevation and outcrop of basal aquifers suggests that the arch was a recharge area in a regional ground-water flow system that extended northward into the present-day Illinois basin. This ground-water regime was not re-

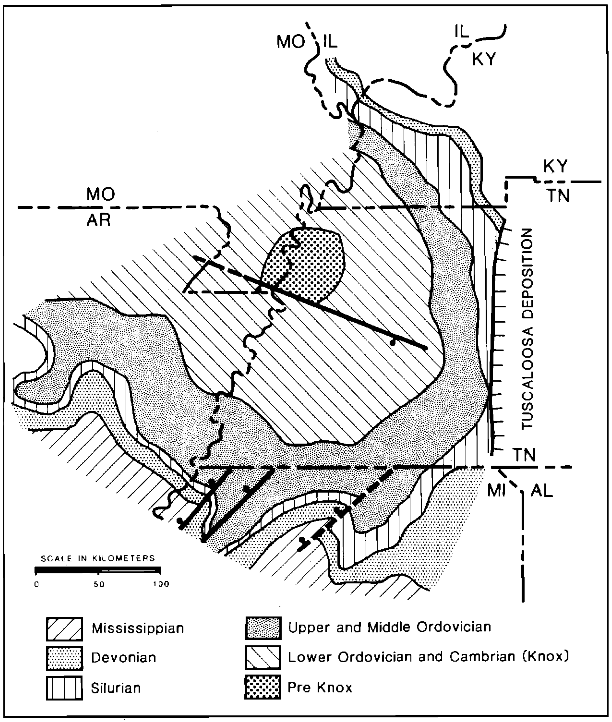

FIG. 8. Pre-Late Cretaceous subcrop map showing erosional surface of Pascola arch and area of Tuscaloosa Formation deposition (after Schwalb, 1982). Upthrown sides of faults are marked.

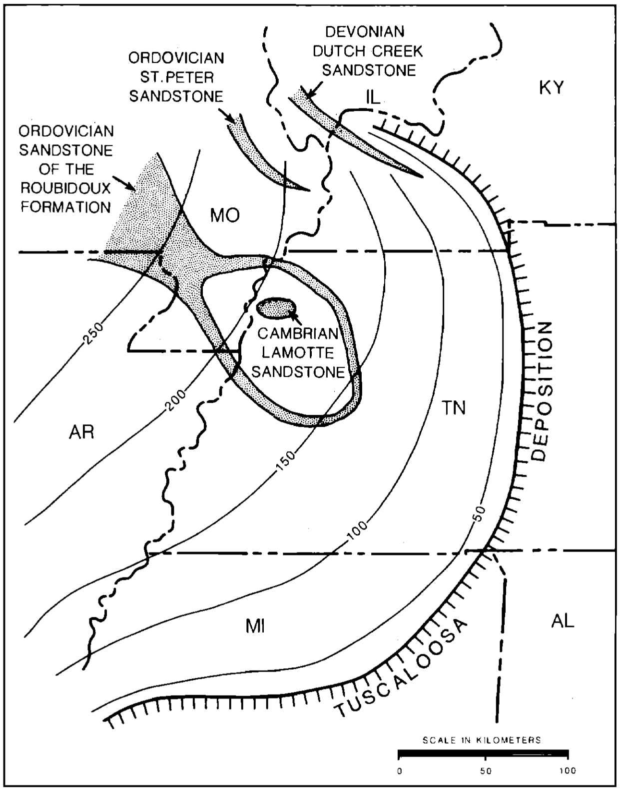

FIG. 9. Reconstruction of minimum elevation on Pascola arch, based on sediment transport into the Cretaceous Tuscaloosa Formation, and subcrop positions of Illinois basin aquifers (after Marcher and Stearns, 1962). Past minimum elevation is contoured in meters above sea level.

lated to current southward and westward flow in the basin resulting from Quaternary incision of the Ohio and Mississippi Rivers (Bond, 1972).

Effects of the uplifted arch on Illinois basin hydrology during post-Early Permian and pre-Late Cretaceous time were numerically modeled. In these calculations, a hydraulic potential gradient was imposed across a basin cross section extending from the Pascola arch northward through the Fairfield basin and onto the Wisconsin arch (Fig. 10), causing south to north gravity-driven flow. In order to err toward underestimating thermal effects of gravity-driven flow, no Permian sediments were included in the cross section.

In addition to uplift in the south, several factors enhance the likelihood of a past flow regime along this south-north cross section. The LaSalle anticlinal belt and Mississippi River arch (Fig. 2) were topographic highs in post-Pennsylvanian time (H. Schwalb, pers. commun.), suggesting that regional slope on the water table was toward the north. These structures, whose trends converge toward the Upper Mississippi Valley district, may have also funneled ground waters into district host rocks. The Precambrian surface in the basin contained south-north-trending valleys which formed large channels in the overlying Mt. Simon aquifer (Schwalb, 1982). Cambrian and Early Ordovician sediment transport patterns resulted in common sandstone facies in the north and more dolomitic facies southward (Ostrom, 1964); many deep sands tend to coarsen northward (Willman et al., 1975), suggesting that permeability increased toward the Wisconsin arch. Finally, fracturing and faulting on the Wisconsin arch (Heyl et al., 1959) may have allowed fluids in deep aquifers to discharge to the surface.

Fluid velocities resulting from uplift of the Pascola arch are difficult to calculate with certainty, because they depend on past values of both aquifer permeabilities and water table elevation on the arch relative to the discharge area. These values can only be estimated. Rather than varying both permeabilities and arch elevation to study the effects of possible flow rates, the calculations presented here arbitrarily assumed a head difference between recharge and discharge areas of 700 m, giving a 1 per mil gradient over the cross section. Uncertainty in flow rates was then accounted for by varying sandstone permeabilities.

Transient calculations made at the onset of this study showed that steady state temperature distributions were achieved quickly with respect to time periods of arch uplift. For this reason, because rates of arch uplift and erosion are difficult to gauge, and due to the extra expense of transient calculations, only steady state results of gravity-driven flow calculations are presented here.

Figure 11 shows results of gravity-driven flow calculations for different basin configurations, sandstone permeabilities, and basement heat flows. Plot (a) gives

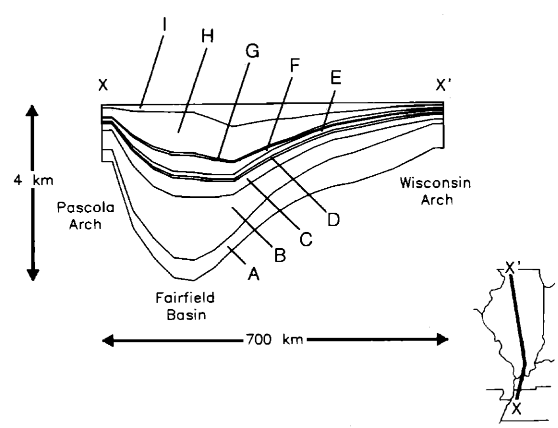

FIG. 10. Illinois basin cross section used in gravity-driven flow calculations. Vertical exaggeration is 85:1. Cross section extends 700 km from the Pascola arch through the Fairfield basin and toward the Upper Mississippi Valley district on the Wisconsin arch.

the resultant flow field and temperature distribution from previously assumed permeabilities (Fig. 5) and a basement heat flow of 1.5 HFU (63 mW/m^2), the continental average (Lee and Uyeda, 1965). Plot (b) shows results of increasing assumed sandstone permeabilities by a factor of 4 relative to plot (a) and allowing discharge through a high permeability zone on the Wisconsin arch. This zone, in which vertical permeability was set to one Darcy, simulates discharge through a broad area of fractured strata. Other conditions remain identical to plot (a). Vertical discharge on the Wisconsin arch might also have resulted from the presence of a past ground-water divide (Freeze and Cherry, 1979, p. 193–195), but paleohydrologic conditions to the north of the Upper Mississippi Valley district are unknown due to extensive erosion of the Paleozoic section there.

Plot (c) demonstrates possible effects of funneling ground waters toward the Upper Mississippi Valley district due to the orientations of the LaSalle anticlinal belt and Mississippi River arch (Fig. 2). Conditions in (c) are the same as those in (b), except that flow converges from the midpoint of the cross section to the northern end by an arbitrary factor of 5:1. Converging flow was obtained by manipulating transmissibility coefficients in the numerical procedure to reflect a radial flow regime in the northern basin.

Plot (d) assumes identical conditions, except that basement heat flow is 2.5 HFU (105 mW/m^2), considerably greater than the continental average. Plot (e) was also calculated for a basinwide average heat flow of 2.5 HFU (105 mW/m^2), but high heat flow was concentrated in the southern basin, as suggested by various geologic data. This calculation assumed a flux of 1.5 HFU (63 mW/m^2), except in the southernmost 250 km of the cross section, where it was 4.28 HFU (179 mW/m^2).

These calculations suggest possible gravity-driven flow velocities of several m/yr, more than three orders of magnitude greater than those predicted for compaction-driven flow. Relatively large velocities result from high permeabilities of basal aquifers (Fig. 12), a decrease in viscosity of pore fluids with depth (Fig. 13), and an additional "apparent" hydraulic head which is produced by thermal expansion of upflowing relative to downflowing fluids (Torrance et al., 1980). The latter effect may be partially offset in nature by increased salinities of upflowing fluids.

High flow rates through the basin allow transport of warm fluids onto the Wisconsin arch, causing rotation of isotherms toward the direction of flow, decreased geothermal gradients to the south, and markedly increased gradients to the north. Similar effects occur in modern basins (Smith and Chapman, 1983; Hitchon, 1984). Figure 14 shows that this thermal regime results from entrainment of basement heat flow by advecting ground waters. Advection in the recharge area and deep basin diverts a portion of the basement heat flow along fluid flow paths before it can be conducted to the surface. As ground waters from the deep basin move toward the recharge area, the fluids lose some of their acquired heat by conduction, causing increased heat flow at the surface relative to the basement. Conduction to the surface, however, may be insufficient to cool ground waters to a normal geotherm, and ascending fluids on the Wisconsin arch can cause an area of anomalous heat flow. This effect becomes more pronounced as increased rates of advective transport overwhelm the ability of conduction to dissipate heat carried from the deep basin.

Temperatures attained in shallow rocks on the Wisconsin arch depend on the presence of permeability structures to allow fluids to wellup, convergence of ground waters toward the arch, basin heat flow, and fluid flow rates. Figure 15 shows temperature profiles with depth on the arch, where lines (a–e) correspond to calculation results in Figure 11. The profile from a calculation in which there was no tendency for upwelling, line (a), shows a slight increase in temperature relative to a normal conductive profile (broken line), but those in which upwelling occurred (b–e) show a greater ability to heat shallow rocks. Calculations in which ground waters converge toward the arch (c–e), as might have occurred due to the orientations of the LaSalle anticlinal belt and the Mississippi River arch, also produce greater arch temperatures because fluids move more quickly through shallower rocks where conductive heat loss to the surface is rapid. This effect is shown by the shift of line (c) relative to line (b).

Basin heat flow exerts a profound effect on Wisconsin arch temperatures. Line (d) in Figure 15 shows effects of increasing basin heat flow from 1.5 to 2.5 HFU (63–105 mW/m^2), under the same conditions assumed for line (c). High heat flow anywhere in the basin was found to affect arch temperatures. Line (e), made assuming heat flow of 1.5 HFU except in the southern 250 km of the cross section, is significantly shifted from the 1.5 HFU (63 mW/m^2) results, line (c), toward the 2.5 HFU (105 mW/m^2) results, (d), despite the 450 km between the high heat flow zone and the arch.

Increased fluid velocities usually cause greater temperatures in areas of ascending flow, but an optimum velocity for given conditions exists, beyond which ascending flows actually decrease in temperature. This effect, also reported by Torrance et al. (1980) and Garven (1985), is shown in Figure 16 for four series of calculations of various basin configurations and heat flows. Under conditions of small velocities, increased flow rates tend to increase temperatures on the Arch greatly by reducing conductive heat loss from migrating fluids. After a point of dimin-

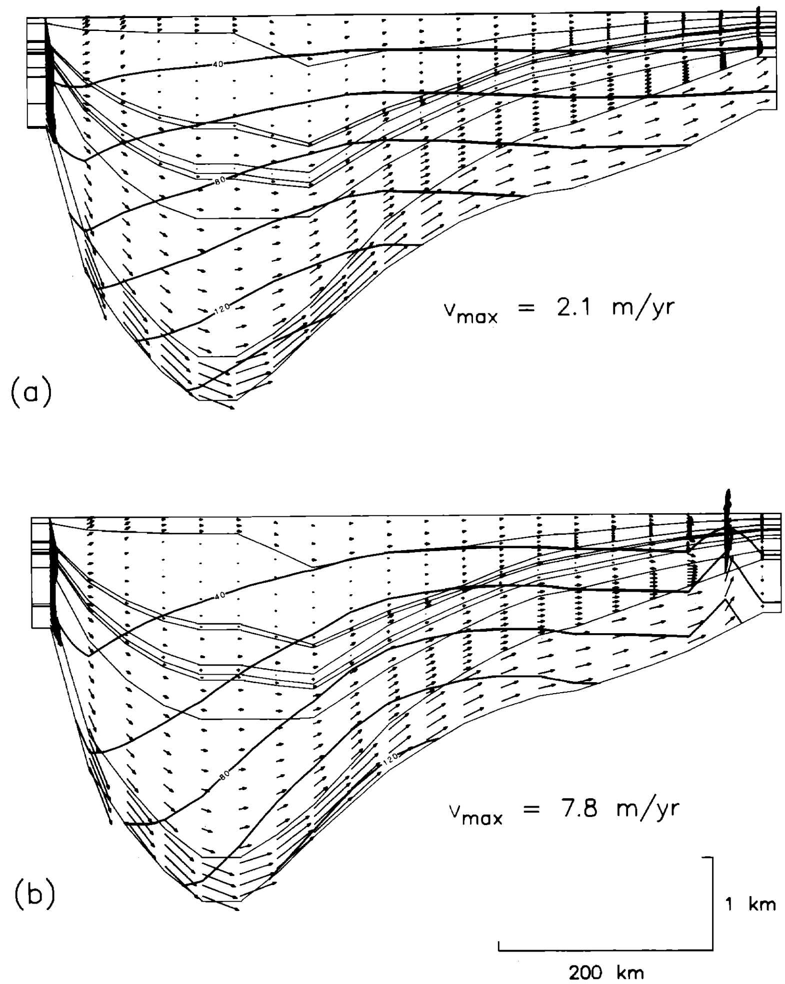

FIG. 11. Results of gravity-driven flow calculations. Arrows represent fluid true velocities, fine lines show contacts between stratigraphic units, and bold lines are isotherms at 20°C intervals. Plot (a) shows expected flow from a 700-m elevation difference across the basin and permeability assumptions in Figure 5, (b) shows the effects of increased velocity and fluid discharge through a high permeability zone, and (c), the effects of a 5:1 converging flow system beginning 350 km from the north end of the cross section. Plot (d) shows the effects of high heat flow, and (e), the results of high heat flow in the southern 250 km of the cross section and normal heat flow to the north.

ishing returns at about 20 m/yr, however, further increase in flow rate only serves to dilute available heat supplied by the basement, because the capacity for advective transport exceeds available heat flow, and arch temperatures decrease.

Discussion

Results of this study show that gravity-driven ground-water flow is more likely than compaction-driven flow to have formed the Upper Mississippi Valley district from Illinois basin fluids. Compaction-driven flow during subsidence of the basin was not effective in transporting heat within the basin because migrating fluids moved too slowly to avoid conductive cooling. The basin was not overpressured during subsidence because of small burial rates, low shale content, and presence of deep aquifers. For this reason, episodic dewatering events are unlikely to have occurred.

Gravity-driven flow due to moderate topographic gradients across the basin, however, could have formed the district by driving warm fluids from the deep basin onto the Wisconsin arch. Factors favoring Mississippi Valley-type genesis by gravity-driven flow across intracratonic basins are: moderate velocities of m/yr through deep aquifers, converging flow toward deposition sites, permeability structures to allow fluids to move toward the surface, and high heat flow in the parent basin.

Results of this study probably also apply to district genesis from other nearby sources, such as the Michigan and Forest City basins. Both basins are tectoni-

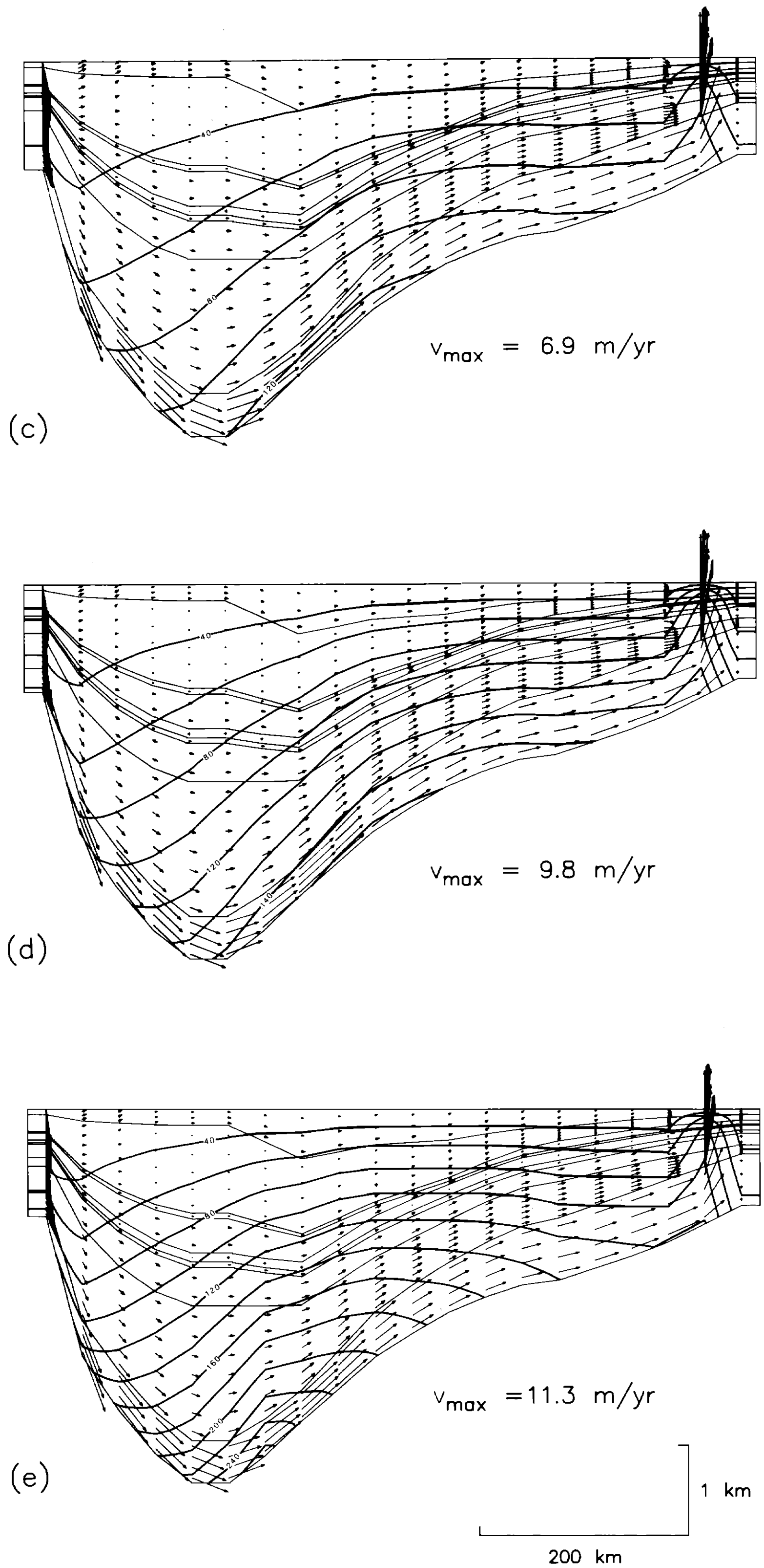

FIG. 11 (cont.). (c), (d), and (e).

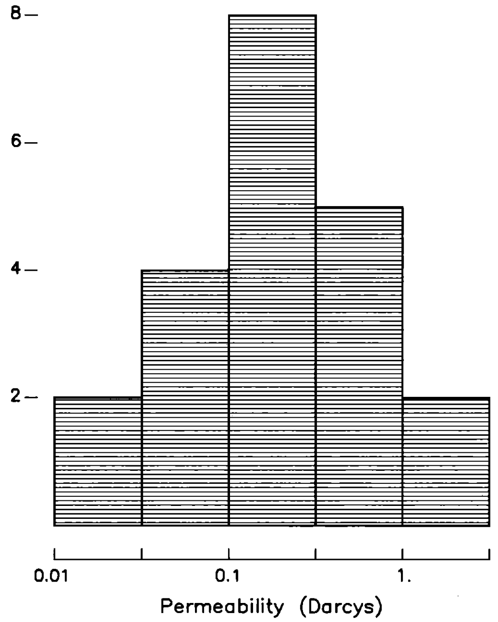

FIG. 12. Histogram of measured permeabilities of deep sandstone aquifers in Illinois basin, from deep gas storage tests (Buschbach and Bond, 1974).

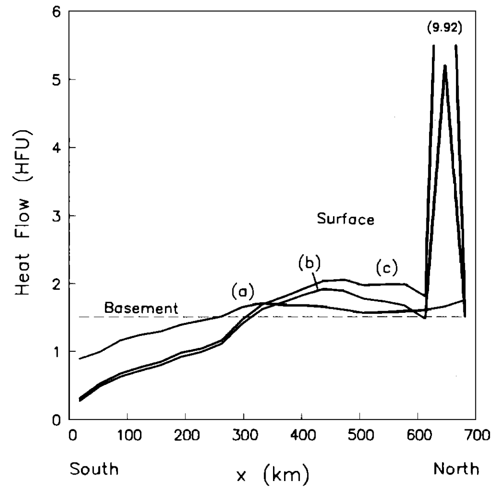

FIG. 14. Basement (broken line) and surface conductive heat fluxes along the cross section for calculations (a–c) in Figure 11.

cally similar to the Illinois basin. The Illinois basin subsided at an average 0.002 cm/yr (McGinnis et al., 1976), the Michigan basin at 0.0024 cm/yr (Fischer, 1975, p. 68), and the shallow Forest City basin more slowly. Compaction-driven flow, however, may be the dominant ore-forming process in other districts associated with shaly, rapidly subsiding basins such as the Gulf Coast or the Ouachita basin. Compaction-driven flow in such basins would be more rapid, overpressuring more common, and gravity-driven flow would be impeded by low permeabilities.

Predicted fluid velocities of m/yr are consistent with interpretations by Lavery and Barnes (1971) and McLimans et al. (1980) that the Upper Mississippi Valley district formed slowly over geologic time periods and with Barnes' (1983) calculation of necessary flow rates through the St. Peter to account for Upper Mississippi Valley mineralization, as well as Roedder's (1976, p. 69–71, 88–89) estimate of velocities through Mississippi Valley-type deposits in general. District origin from gravity-driven flow due to post-Early Permian–pre-Late Cretaceous uplift of the Pascola arch further agrees with McLimans' (1977) estimate of a Pennsylvanian through Cretaceous bracket on the age of mineralization and with fission-track dating of hydrothermal alteration in basement rocks

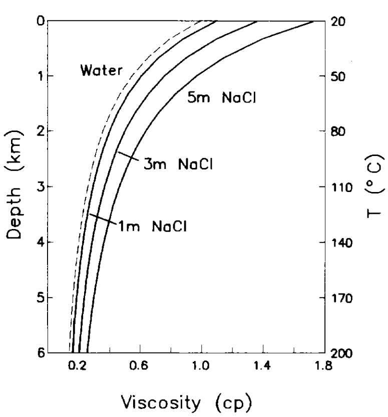

FIG. 13. Viscosities of water (broken line) and 1, 3, and 5 molal NaCl solutions with depth, assuming 30°C/km geothermal gradient and hydrostatic pore pressure (data of Phillips et al., 1980). Viscosities decrease sharply with depth, regardless of salinity.

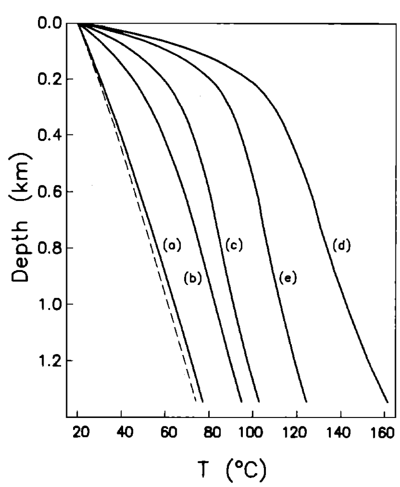

FIG. 15. Temperature profiles with depth within discharge area for calculations (a–e) in Figure 11.

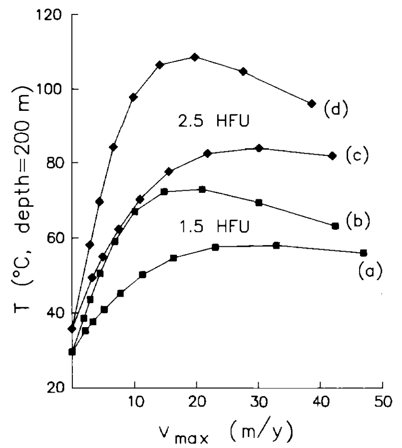

FIG. 16. Temperature at the 200-m depth in the discharge area vs. maximum fluid velocity for four series of calculations, (a–d). Series (a) and (b) were made for basement heat flows of 1.5 HFU (63 mW/m^2), (c) and (d) at 2.5 HFU (105 mW/m^2). Series (a) and (c) did not consider converging flow toward the north, but flow converged 5:1 for the northern 350 km of the cross section in (b) and (d).

(Zimmerman, 1981; Shieh, 1983). Isotopic evidence that predominantly meteoric waters make up present Illinois basin deep brines (Clayton et al., 1966) also strongly supports the possibility of a past gravity-driven flow regime.

Temperatures which could have been produced on the Wisconsin arch by gravity-driven flow depend on past heat flow into the Illinois basin. Temperatures of 75° to 120°C reported in early fluid inclusion studies (Newhouse, 1933; Bailey and Cameron, 1951) agree with calculations made assuming heat flows of 1.5 to 2.5 HFU (63–105 mW/m^2; see Fig. 15). McLimans (1977), however, reported some higher temperature estimates from district fluid inclusions and 5 percent of his data is in the range 160° to 220°C. Gravity-driven flow would have required abnormal heat flows in the basin to produce these temperatures on the arch. Such an interval of high heat flow while the Pascola arch was uplifted is consistent with previously mentioned geologic data from the southern basin.

Banding in Upper Mississippi Valley ores (McLimans et al., 1980), which has been attributed to effects of episodic fluid discharges (Cathles and Smith, 1983), can also be caused by cyclic variation in gravity-driven flow. Such variation could be caused by meteorologic precipitation cycles at the recharge or discharge area, tectonic activity in the recharge area, or permeability changes in the deposition site due to dissolution of host rocks and precipitation of ore and gangue minerals. Figure 16 shows that small variations in ground-water velocity can cause significant changes in temperature in the discharge area which, in turn, could lead to chemical changes at deposition sites.

These calculations also explain the apparent inconsistency of largely uncemented aquifers with silicified veins and fractures in the Upper Mississippi Valley district (Ludvigson et al., 1983). Figure 11 shows that expected temperature gradients along the basal aquifers can be extremely small, much less than a °C/km, while vertical gradients are tens and hundreds of °C/km (Fig. 15). To the extent that silica solubility at estimated pH values for Mississippi Valley-type ore fluids (Giordano and Barnes, 1981) and at low pressures is primarily controlled by temperature (Fournier and Rowe, 1966), relatively little silica should be found in deep sands. Thicknesses of these aquifers and relatively large surface areas of sands in general (Leamnson et al., 1969) would allow precipitated silica to be dispersed in minute quantities. Ascending fluids channeled in a fracture or vein, however, would undergo a rapid temperature drop and precipitate silica within a restricted area.

An origin of the Upper Mississippi Valley district by gravity-driven flow does not offer a simple explanation of the source of salts in ore solutions. The first fluids driven from the basin would be connate waters, which might have been able to supply sufficient metals to form the district (see Carpenter et al., 1974). There is some evidence that district fluids became less saline with time (McLimans, 1977, p. 75–87), but it is not clear that enough connate fluids were available to preheat rocks along the flow paths to deposition sites hundreds of km from the deep basin. If not, meteoric waters that entered aquifers on the Pascola arch must have become saline by dissolving evaporites along the 700-km traverse to the Wisconsin arch.

Although gypsum and anhydrite are exploited from Mississippian evaporites (McGregor, 1954; Saxby and Lamar, 1957; Harvey, 1964; McGrain and Helton, 1964) and anhydrite is commonly observed in Ordovician and Cambrian drill cuttings, the only evidence of halide occurrence in the basin are occasional cubic casts (I. Diaby, pers. commun.; H. Schwalb, pers. commun.). A possible explanation is that evaporites in the deep basin have not been discovered there due to very limited petroleum exploration in rocks older than Devonian (Brehm, 1971) or that original halides have been depleted through leaching by ground waters. The Michigan basin contains 12 percent evaporites by volume, including thick Silurian salt deposits (Ells, 1971; King, 1977, p. 27–35), but there is no hydrologic evidence linking this basin to Upper Mississippi Valley mineralization.

Finally, study results can be applied to exploration strategies for Mississippi Valley-type deposits in continental interiors. Exploration targets should be located in likely discharge areas of past or present regional ground-water flow regimes. As such, strategies should take into account not only the presence of possible parent basins, but elevations and tectonic histories of basin margins distant from targets. Results of this study show that deposit genesis by gravity-driven flow requires parent basins with topographic head gradients and aquifer permeabilities adequate to allow ground-water flow rates of approximately m/yr. Structural or stratigraphic features which cause ground waters to converge or ascend also help localize deposits.

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January 3, May 16, 1985

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APPENDIX

Mathematical Model

Compaction-driven flow simulations in this study employed a mathematical model of fluid flow and heat transfer in a deforming porous medium. Gravity-driven flow simulations were made with the same model, for the special case of no subsidence and constant porosity. The model is summarized here and treated in detail by Bethke (1985b).

Model calculations are performed by cyclically solving finite difference approximations to three differential equations describing continuity of the porous medium, fluid flow within the medium, and heat transfer by conduction and fluid advection. The differential equations are derived in a Lagrangian reference frame, which remains fixed with respect to subsiding sediments but moves relative to fixed elevation. The equations are also derived in general curvilinear coordinates (Mastin, 1982; Thompson, 1982) in order to allow modeling of flow in basins with irregular geometries. Equivalent equations derived in Cartesian coordinates are presented in this summary, because the Cartesian equations are simpler and more familiar than their curvilinear counterparts.

Medium continuity equation

An equation describing continuity of the medium is necessary, because strata at different depths in a compacting basin subside at different velocities, much as coils of a spring move at different speeds as the spring is compressed. The continuity equation is given by:

$$\frac{\partial}{\partial z} v_{zm} = \frac{1}{(1-\phi)} \frac{\partial \phi}{\partial t},$$

where z is positive downward, v_{zm} is the subsidence velocity of the medium, and ϕ is porosity. Porosity is typically defined as a function of depth, as in this study. In simulations in which pore pressures approach lithostatic, however, porosity is more appropriately taken as a function of effective stress (Jaeger and Cook, 1976, p. 219). Solution of this equation over the domain gives consistent values for v_{zm} and $\partial\phi/\partial t$, which are used in solving the fluid flow equation.

Fluid flow equation

Flow within the medium is assumed to be driven by variation in a hydraulic potential function, which is the product of Hubbert's (1940) potential and fluid density:

$$\Phi = P - \rho gz.$$

P is fluid pressure, ρ is density, and g is the acceleration of gravity. This function, which has units of pressure, is convenient in compaction-driven flow problems because it allows simultaneous consideration of effects of both pressure and elevation changes on flow. While no potential function gives a totally rigorous description of flow of a heterogeneous fluid (Bear, 1972, p. 641–655), Darcy's law, written in terms of hydraulic potential for a slightly compressible and expansive fluid in a regional flow system, is a good approximation to the general momentum equation.

The fluid flow equation is derived from an equation of state for a fluid of constant composition:

$$\frac{1}{\rho}\,\partial\rho = \beta\partial P - \alpha\partial T$$

(Lewis and Randall, 1961, p. 25; Domenico and Palciauskas, 1979), where β and α are coefficients of compressibility and expansivity and T is temperature. This equation can be expanded by the relations:

$$\partial\rho = \frac{1}{V}\,\partial m - \frac{\rho}{V}\,\partial V,$$

$$\partial V = \frac{V_b}{(1-\phi)}\,\partial\phi,$$

and

$$\partial P \simeq \partial\Phi + \rho g\partial z.$$

Variables m and V are the fluid mass and volume within a bulk volume, V_b. Combining these equations with Darcy's law gives the fluid flow equation:

$$\phi\beta\left(\frac{\partial\Phi}{\partial t} + \rho g v_{zm}\right) = \frac{1}{\rho}\left[\frac{\partial}{\partial x}\left(\frac{\rho k_x}{\mu}\frac{\partial\Phi}{\partial x}\right) + \frac{\partial}{\partial z}\left(\frac{\rho k_z}{\mu}\frac{\partial\Phi}{\partial z}\right)\right] - \frac{1}{(1-\phi)}\frac{\partial\phi}{\partial t} + \phi\alpha\frac{\partial T}{\partial t}.$$

Here, k is permeability and μ is viscosity. From left to right, terms in this equation give the change in hydraulic potential due to loss of potential energy, divergence of Darcy fluxes, pore volume collapse, and thermal expansion of the pore fluid.

Heat transfer equation

Heat transfer by conduction and advecting ground waters is described by:

$$[\rho\phi C_w + \rho_r(1-\phi)C_r]\frac{\partial T}{\partial t} = \nabla\cdot(K\nabla T) - \nabla\cdot(\rho h_w q) - \frac{\rho h_w}{(1-\phi)}\frac{\partial\phi}{\partial t},$$

where $\nabla\cdot$ and ∇ are the divergence and gradient operators, C_w and C_r are heat capacities of fluid and rock grains, ρ_r is the density of rock grains, K is thermal conductivity, h_w is fluid enthalpy, and q is the specific discharge vector. This relation is the advection-conduction equation (Stallman, 1963) with an extra term describing the heat content of fluid lost from sediments during compaction.