1988


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Aspects of subsurface brine formation in passive continental margin evaporite basins

Ranganathan, Vishnu, Ph.D.
The Louisiana State University and Agricultural and Mechanical Col., 1988
ASPECTS OF SUBSURFACE BRINE FORMATION IN
PASSIVE CONTINENTAL MARGIN EVAPORITE BASINS

A Dissertation
Submitted to the Graduate Faculty of
Louisiana State University and
Agricultural and Mechanical College
in partial fulfillment of the
requirements for the degree of
Doctor of Philosophy
in
The Department of Geology and Geophysics

by
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December 1988
I wish to thank Jeff Hanor, who stimulated my interest in the topic of subsurface brine formation in evaporite basins, was a source of much encouragement throughout the course of this dissertation, and showed enthusiasm in this work through constructive criticisms and discussion. The seeds of this work sprung as a consequence of an exciting course taught by Jeff in the Spring of 1985 an outcome of which was the first of three papers comprising this dissertation (Chapter 2).

During the course of this work, my colleague Dave Evans was extremely helpful and supportive. Dave shared many of his ideas concerning thermohaline convection in sedimentary basins with me and kindly allowed me to use an interactive graphics program that he developed for the USGS code SUTRA. For this, I wish to thank Dave.

To Tissa Illangasekare who taught me two courses in groundwater hydrology at LSU and showed me an engineer's perspective of groundwater hydrology, I am grateful. I wish to thank Dean Adrian, Richard Condrey, Chad McCabe, Jeff Nunn and Rex Pilger for serving as members of my dissertation committee and for their reviews of this dissertation. During my first two years at LSU, Cathy Duncan and Dave Wilensky helped me become familiar with the use of the VAX minicomputer. At various stages of this project, Mary Lee Eggart and John Duplechin drafted figures and prepared illustrations for me, and I wish to express my appreciation of their help.
Finally, I wish to thank two people who may never read this dissertation, but without whose support this whole project would have been very, very difficult - my father and mother.
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This dissertation consists of three papers which quantitatively model subsurface brine formation in passive, continental margin evaporite basins. In the first paper, a one dimensional finite difference model is developed to evaluate the brine-forming role of molecular diffusion of dissolved NaCl in the subsidence history of a hypothetical sedimentary basin containing a thick basal halite bed. It is shown that diffusion of dissolved NaCl from evaporites may be an important brine-forming process in young passive margin evaporite basins where evaporites are flat lying or gently dipping and that vertical fluxes of dissolved NaCl are great for basins with slow subsidence rates and sandy sediments.

Whereas diffusion may be an important process in areas with a thick, relatively horizontal halite bed, solute-induced fluid density variations near salt domes may give rise to density driven groundwater flow. In the second paper, it is shown that solute-induced fluid density gradients can induce convective flows with maximum velocities of $10^{-2}$ to $1$ m/yr even in low-permeability sediments, brine plumes can form very rapidly by solute-induced convection, and up to $2$ km$^3$ halite can be dissolved in $1$ Ma.

In the third paper, the combined effects of geopressure and solute-induced fluid density gradients are modeled around a hypothetical salt dome bearing some resemblance to Welsh Dome in South
Louisiana where a perched brine plume has been shown to occur above the dome, and near a fault. It is shown that geopressured fluids escaping up a fault zone on the flank of the 'Welsh-type' salt dome may create a time-transient perched brine plume above the dome in very short times, $10^{-1}$ Ma, provided that permeability of the geopressured zone is sufficiently high. However, enormous volumes of water must be sourced from the geopressured zone to form the perched brine plumes, volumes which appear to be very large in comparison to estimates of water available upon compaction of geopressured sediment.
CHAPTER 1

INTRODUCTION
1.1 Continental Rifting and Evaporite Basins

Evaporite basins occur in every system from the Precambrian to the Cenozoic (Halbouty, 1979) and are found in many parts of the world (Fig. 1.1). During the early stages of formation of the Atlantic Ocean in the Late Triassic and Early Jurassic, several rift basins formed on the eastern continental margin of North and South America and the western continental margin of Africa. In some of these basins, sequences of massive salt were deposited, more than 1 km thick (Ayme, 1965; Belmonte et al., 1965; Demaison, 1965; Burke, 1975; Halbouty 1979), a consequence of influx of seawater over shallow sills into the rift basins and evaporation due to the arid climate (Burke, 1975; Hsu, 1972). In the early 1970’s deep sea drilling revealed that much of the Mediterranean is underlain by extensive salt deposits of Miocene age. The Dead Sea and Red Sea rift basins have also been found to have thick evaporites (Neev and Emery, 1975; Heybroek, 1965). Indeed, thick evaporite sequences are considered to be a characteristic feature in the early history of many basins which underwent rifting in arid environments. A common feature of salt basins on the eastern continental margin of the Americas and the western continental margin of Africa is that thick sequences of nonevaporitic sediments were deposited on top of the evaporites in Cretaceous and Tertiary time.

The Gulf of Mexico was one such rift which opened from about 180 Ma to 130 Ma ago in the same direction as the central North Atlantic, and although the geometry of early rifting is uncertain, according to one
Fig. 1.1 Major salt basins of the world (modified after Halbouty, 1979, p. 13, 14). Five separate salt dome basins occur within a larger Gulf Region salt basin in the U.S. part of the Gulf of Mexico. Dark lines indicate possible extent of relatively unexplored salt basins.
model, sea floor spreading was accomplished by movement along left-lateral transform faults (Pilger, 1981). In the northern part of the U.S. Gulf Coast, South Arkansas, where the post-evaporite sedimentary sequence is relatively thin, the salt is either bedded or in the form of broad swells or salt pillows (Moore and Druckman, 1981, p. 599, p. 602). Where thick sedimentary sequences have been deposited on top of the evaporites, salt diapirs rise through sediments 5 to 6 km above the parent salt in North Louisiana (Lobao and Pilger, 1985) and more than 10 km in South Louisiana (Buffler and Sawyer, 1985, p.336).

1.2 Subsurface Brine Formation in Evaporite Basins

Not surprisingly, subsurface brines are common in many evaporite basins. Chloride content in pore waters of deep-sea sediments above evaporites in the Mediterranean increases with depth from 0.6 moles/l at the sediment-seawater interface to 5 moles/l above the evaporites and has been interpreted as being diffusion-controlled (Fig. 1.2; McDuff et al., 1978). Ten DSDP (Deep Sea Drilling Project) sites in the Mediterranean gave evidence of the existence of halite-containing sediments beneath the underlying seabed in the form of continuously increasing Na and Cl enrichment in interstitial waters with depth (Sayles et al. 1973). Diffusion-controlled salinity profiles also occur in pore waters of sediments above Miocene evaporites in the Red Sea (Fig. 1.3; Whitmarsh, et al., 1974; Manheim, et al., 1974). Pore water salinity profiles in the Timor Trough, off the coast of Northwest Australia also appear to be diffusion controlled, judging by the borehole data of Cook (1974), although no evaporites
Fig. 1.2 Chloride content of pore waters in sediments above Miocene evaporites at two Deep Sea Drilling Project sites in the Mediterranean (after McDuff et al., 1978). Water analyses are represented by crosses. The lines are Cl profiles calculated assuming diffusion of Cl from evaporites to the sediment-water interface.
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were encountered at this drill site. Gieskes et al. (1980) noted the occurrence of saline brines with as much as 1 mole/l Cl (almost twice as much as seawater) in pore waters of sediments off the Northwest African continental margin where evaporites are known to occur, although at this location, they suggested that some lateral brine migration occurred through aquifers, rather than by diffusion alone.

Pore waters have salinities of up to 200 g/l TDS in the Alberta Basin and up to 425 g/l TDS in the Michigan Basin (Hanor, 1979). Pore waters below salt deposits in the Gabon Basin of West Africa have almost 300 g/l TDS (Cassan et al., 1981, p. 126). Subsurface brines with more than 100 g/l total dissolved solids (TDS) are widespread at depths of 2 to 4 km in South Louisiana (Hanor et al., 1986), and chemical analyses show that pore waters around some salt domes have more than 300 g/l TDS. The occurrence of subsurface brines spatially related to evaporites has been observed along the Atlantic Coastal Plain, in a transect from New York to Florida (Manheim and Horn, 1968), on a scale of tens of kilometers in South Arkansas (Hanor, 1984a), and on a scale of hundreds of kilometers in South Louisiana (Fig. 1.4; Hanor, 1984b).

In addition to the regional occurrence of subsurface brines on a scale of tens to hundreds of kilometers, there exist smaller scale anomalies (1 km), near the tops of salt domes, where salinities are locally higher than away from the domes (Manheim and Bischoff, 1969; Hanor and Bailey, 1983; Bennett and Hanor, 1987). The anomalies have been attributed to be due to diffusion of dissolved NaCl (Manheim and
Fig. 1.4 Salinities of groundwaters at depths of 1000 ft (300 m), 2000 ft (600 m), 3000 ft (900 m), 4000 ft (1200 m) in South Louisiana (after Hanor, 1984a). Contours are in g/l.
Bischoff, 1969), groundwater flow induced by fluid density gradients (Hanor, 1987a), and escape of fluids up faults from the geopressed zone (Bennett and Hanor, 1987).

Recent regional hydrodynamic modeling studies of the U.S. Gulf Coast have been concerned mainly with compaction flow and geopressure development and not with transport of dissolved salt and brine formation (Keith and Rimstidt, 1985; Bethke, 1986; Blanchard, 1987; Bethke et al., 1988). Maximum fluid velocities due to compaction flow are low, on the order of $10^{-3}$ to $10^{-2}$ m/yr (Blanchard, 1987; Bethke et al., 1988).

The objectives of this study are to quantify rates at which subsurface brine formation may occur in evaporite basins, using numerical modeling. Chapter 2 of this dissertation deals with the early stages of passive margin evaporite basins where sediments are horizontally bedded and vertical diffusion is likely to be the dominant transport process. In contrast, Chapter 3 and 4 deal with brine formation around older passive margin evaporite basins where salt diapirs exist. In contrast to diffusion of NaCl dissolved from halite which is of great importance in the early stages of passive margin evaporite basins, brine formation around salt diapirs is likely to occur by advection and mechanical dispersion due to density-driven groundwater flow and flow of fluids escaping the geopressed zone.
There are several practical reasons for studying the possible hydrodynamics of subsurface brine formation in salt dome basins. For example, the primary area of oil production at Welsh Dome in South Louisiana corresponds approximately to a plume of anomalously saline water extending away from the dome to the northeast (Bennett and Hanor, 1987). Sulphide mineralization associated with many Gulf Coast salt domes is closely related to cap-rock formation (Kyle and Price, 1986) and cap rock formation is closely related to the dynamics of subsurface salt dissolution. Cementation in sand aquifers has been documented to be more pervasive near salt domes, albeit in certain stratigraphic intervals (Stuart, 1970, p. 62; McManus and Hanor, 1988), and the cementation patterns are postulated to have resulted from brine density flow near the salt dome (Stuart, 1970, p. 62).

Vast quantities of liquid chemical waste are injected into the subsurface of South Louisiana, and some of the injected wastes are fortuitously located near or between salt domes. With the heavy usage of groundwater drawn from Coastal Plain sediments in the Gulf Coast, it behooves us to gain a better understanding of the possible dynamics of groundwater flow in not only the shallow surficial aquifers, but also the deep aquifers. Although this dissertation does not discuss the effects of anthropogenic perturbations which may affect the quality of waters in aquifers surrounding salt domes in the Gulf Coast, it is of importance to study processes that have operated over geologic time scales in order to make sensible decisions concerning groundwater use even over the short time interval of our lives.
1.3 The Origin of Subsurface Brines

Most controversy concerning the origin of subsurface brines centers around three main hypotheses: subsurface dissolution of halite (Land and Prezbindowski, 1981; Hanor et al., 1986), burial of subaerially produced brines (Carpenter and Trout, 1978), and membrane filtration (Russell, 1933; Bredehoeft et al., 1963; Kharaka and Berry, 1974; Graf, 1982). The development of thought on the origin of subsurface brines has been recently reviewed by Hanor (1987b).

The evidence in favor of subsurface dissolution of halite is simple, the spatial relationship between salty waters and evaporites, first documented by Hitchcock (1845, p. 188 - 189). High Br/Cl ratios relative to seawater are cited as evidence, although not indisputably so, in favor of burial of subaerially produced hypersaline brines because Br is not easily incorporated into halite and is concentrated in seawater (Carpenter and Trout, 1978). An alternative explanation for high Br/Cl ratios is the dissolution of evaporite minerals containing Br. Because these minerals are the last to precipitate upon evaporation of seawater, they would likely be the first to dissolve.

The evidence in favor of membrane filtration is tenuous at best and ignores what is known about the hydrodynamics of sedimentary basins. For example, in laboratory experiments, semipermeable membrane effects were observed for compacted clay discs under fluid
pressure gradients of 270 bars/cm (Coplen and Hanshaw, 1973, p. 2306). The pressure gradient in this experiment was equal to a hydraulic head gradient of $2.7 \times 10^3$, which is $5 \times 10^3$ times the hydraulic head gradient in natural geopressured sediments of the Louisiana Gulf Coast (approximately 5, according to the data of Bennett, 1985).

1.4 Aims of this study.

The increase in Cl content of pore waters above Miocene evaporites in the Mediterranean and the Red Sea has been explained by diffusive transport of dissolved salt from the evaporite to the sediment-water interface (McDuff et al., 1978; Manheim, et al., 1974). In Chapter 2, a finite difference numerical model is developed to evaluate the brine-forming role of one-dimensional vertical diffusive transport of dissolved NaCl in a subsiding sedimentary basin with a basal evaporite layer. It is shown that for basins with slow subsidence rates and sand-dominated sediments, diffusion of NaCl may be an important brine-forming process. However, the deviation of diffusion-generated salinity profiles from observed salinity profiles in some older evaporite basins indicates that some degree of upward advection of pore waters may occur along with mechanical dispersion.

While diffusion appears to an important brine forming process where sediments are horizontal or dip very slightly, as in the case of deep sea sediments from the Mediterranean and Red Sea, it is not likely to be of great importance in brine formation around salt domes. Instead,
advection and mechanical dispersion are likely to be the dominant processes near salt domes because of gravitational instability in the pore water column.

In Chapter 3, an evaluation is made of the importance of density-driven groundwater flow around salt domes which arises as a consequence of solute concentration gradients or temperature gradients. Using a U.S. Geological Survey numerical model for variable density groundwater flow, SUTRA (Voss, 1984), it is shown that brine density flow may be an important process near salt domes where active halite dissolution occurs, with average linear groundwater flow velocities of $10^{-2}$ to 1 m/yr even in low-permeability sediments. Brine plumes can theoretically form quickly, in less than 1 Ma, and salt dome dissolution can be as great as 2 km$^3$/Ma. It is shown that thermal convection is likely to be secondary in importance to brine density flow, but that the effect of geopressed fluids bleeding up into the hydropressed section may be important.

In addition to subsurface brine formation on a scale of tens to hundreds of kilometers, there are small scale anomalies (1 km) associated with the Welsh and Mallard Bay salt domes in South Louisiana (Fig. 1.5; Hanor and Bailey, 1983). Upon more detailed mapping, the salinity anomaly above Welsh Dome was found to be spatially associated with the top of a fault possibly extending into the geopressed zone (Bennett and Hanor, 1987). Bennett and Hanor (1987) postulated that focused flow of geopressed fluids up the fault on
Fig. 1.5 Salinity field near two South Louisiana salt domes (after Hanor and Bailey, 1983). High salinities in wells 4 and 8 correspond to Welsh and Mallard Bay salt domes respectively.
the flank of Welsh may have been partly responsible for the salinity anomaly at Welsh. In Chapter 4, the effect of a short \((10^{-1} \text{ Ma})\) pulse of geopressured fluid flow up a fault flanking a 'Welsh-type' dome is evaluated. It is shown that brine anomalies of the type mapped by Bennett and Hanor (1987) can indeed form in very short times, but that effluxes required from the geopressured zone are extremely large. Two competing factors are shown to determine whether or not a perched brine plume will form: 1) density-driven groundwater flow which will tend to make a perched brine plume sink off the flank of the dome, and 2) upward flow of geopressured fluids through the fault zone. The permeability of the geopressured zone is shown to be a strong control on which of these two processes prevails.

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CHAPTER 2

A Numerical Model for the Formation of Saline Waters due to Diffusion of Dissolved NaCl in Subsiding Sedimentary Basins with Evaporites

V. Ranganathan  J.S. Hanor

(A slightly modified reprint from J. Hydrol., 92: 97 - 120)

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Numerical modeling by a finite difference method incorporating the effects of vertical variations in pressure, temperature, salinity, and porosity shows that mass transport by diffusion may be a major brine forming process in sedimentary basins containing halite beds. Vertical fluxes of dissolved NaCl are greatest when the subsidence rate is slow. At subsidence rates of 20 m/Ma, more than 250 m of halite can be dissolved over 150 Ma. At rapid subsidence rates, such as 500 m/Ma, pore waters are advected downward so rapidly that although a diffusion front develops above the evaporite, this front moves downward as the evaporite subsides, and diffusive fluxes at shallow depths are thus low. However, under such conditions, the pore water column becomes gravitationally unstable as a result of thermally induced density inversions and the potential for convective overturn is enhanced with accompanying advective salt flux upward. Although dissolution and diffusion of NaCl are important processes in the burial history of evaporites, they are not adequate by themselves to produce the regional salinity versus depth profiles that are actually observed in some sedimentary basins. Diffusion-induced gradients progressively increase with depth whereas gradients actually observed in sedimentary basins tend to decrease with depth. The difference is presumably due to the activity of other mass transport mechanisms such as advection, convection, and dispersion, and possibly to chemical reactions.
INTRODUCTION

Analyses of oil field waters show that pore water salinity increases with depth in many sedimentary basins (Campbell, 1929; Meents et al., 1952; White, 1965). Dickey (1966, 1969) pointed out that there is a nearly linear increase in total dissolved solids (TDS) content of oil field waters with depth in areas of northern Louisiana and southern Arkansas. Similar relationships exist in portions of east-central Louisiana and central Mississippi as can be deduced from the data of Hawkins et al. (1963), Collins (1970), and Carpenter and Pickett (1974). These linear gradients are remarkable in that some extend vertically over depths exceeding 2700 m (Fig. 2.1). Hanor (1984) found that TDS increases linearly mainly because the concentrations of dissolved chloride and dissolved sodium, the two most abundant dissolved species on a mass basis, increase in a linear fashion with depth.

Maximum values of pore water salinity also increase with depth in the Alberta, Illinois and Michigan Basins (Fig. 2.1), but the salinity gradient is not constant in these basins. Maximum salinities increase in a curvilinear manner and reach asymptotic values at depths of 2 -3 km.

Hanor (1984) interpreted the linear increase in NaCl with depth in North Louisiana as indicating that steady-state diffusion of NaCl is occurring from evaporites deeper in the section. The average depth of the Louann Salt in North Louisiana is 3000 m and the NaCl concentration extrapolated to this depth, 412 g l⁻¹, is well in accordance
Fig. 2.1. Maximum salinities of pore waters from the Alberta, Illinois, and Michigan Basins, and from the Louisiana Gulf Coast (after Hanor, 1979).
with the solubility of halite expected at this depth. Low dissolved salt concentrations, almost 0 g l⁻¹, are maintained from the surface to a depth of 360 m by active recharge of meteoric waters.

Three hypotheses are in vogue for the origin of subsurface brines. These are: membrane filtration (Bredehoeft et al., 1963; Karaka and Berry, 1974); burial of subaerially produced brines (Carpenter and Trout, 1978); and subsurface dissolution of evaporites (Land and Prezbindowski, 1981).

Some workers have suggested that the chemical composition of pore waters in sedimentary basins may be affected by diffusive mass transport. Richardson (1917) was perhaps the first to suggest that upward diffusion of dissolved NaCl above buried evaporites can significantly raise the NaCl concentration of pore waters within geologically reasonable time. Klimenko (1974) proposed that diffusion of hydrocarbon gases from oil and gas reservoirs raises the concentration of these gases in pore waters and suggested that these dissolved gases can be used as proximity indicators for oil. Manheim and Bischoff (1969) found that pore water salinity systematically increased with depth in sediments above some salt domes in the Gulf Coast and ascribed this to diffusion of dissolved salt. Similar observations were made by Manheim and Paul (1980) in sediments of the Atlantic Coastal Plain which overlie evaporites. According to Kaufmann et al. (1984), isotopic variations of ³⁷Cl and ³⁵Cl occur in pore waters of some oil field brines as a function of distance from evaporites and these isotopic variations are best
explained by diffusive mass transport of $^{37}\text{Cl}$ and $^{35}\text{Cl}$ (Kaufmann et al., in prep.).

The purpose of this paper is to investigate in a systematic way the potential of large-scale molecular diffusion as a brine-forming process in sedimentary basins. We trace the changes in NaCl concentration with depth and time due to diffusion in pore waters in a hypothetical, actively subsiding sedimentary basin with a basal evaporite sequence. We show, using a finite difference method to calculate transient NaCl concentrations and fluxes at different depths, that mass transport by diffusion can be a significant control on salinity gradients in many sedimentary basins which contain evaporites. We show that for a thickening sequence of sand-rich sediments deposited above a subsiding evaporite, for subsidence rates such as 20 m/Ma diffusive fluxes are so large that more than 250 m of halite can be dissolved after 150 Ma. Diffusive mass fluxes are high even for a shale-rich section, provided that subsidence is slow. We also evaluate the gravitational stability of the pore water column as a function of subsidence rate and time and show that density inversion in pore waters occurs only when subsidence rates are extremely high, such as 500 m/Ma. Finally, we examine the modifying effects of upward advection on the salinity profile for the special case of a steady-state salinity profile.

The geologic situation we consider is schematically illustrated in Fig. 2.2. A halite layer initially at a very shallow depth subsides at a constant rate to a maximum depth of 3 km while a thickening sediment
Fig. 2.2. Schematic diagram illustrating an actively subsiding sedimentary basin in the vertical dimension. The base of the sedimentary basin coincides with the top of the evaporite. The porosity of the sediments at the surface is $\sigma_0$. Possible end members of salinity profiles are shown in dashed and dotted lines in the lower half of the figure.
column forms over it. As sediments over the halite bed compact, their porosity decreases exponentially with depth. Two geologic constraints are imposed: (1) at the top of the halite bed, pore waters are always saturated with respect to NaCl; and (2) sedimentation keeps up with subsidence such that the sediment water interface stays fixed at a datum whose elevation is zero, and the sediments being deposited at the surface initially contain fresh pore water. Given these conditions, how will the salinity profiles of pore waters change as the evaporite subsides?

MATHEMATICAL FORMULATION OF THE PROBLEM

The diffusion equation

The equation governing mass transport by molecular diffusion alone in a porous sediment in one dimension is:

\[
\frac{\partial}{\partial t} \{C(x,t)\} = \frac{D(x,t)}{\rho} \frac{\partial}{\partial x}\{D(x,t) \frac{\partial}{\partial x}[C(x,t)]\}
\]

subject to:

\[
\begin{align*}
C(x,0) &= 0 \text{ g l}^{-1}\text{NaCl,} \\
C(0,t) &= 0 \text{ g l}^{-1}\text{NaCl,} \\
C(L,t) &= s(T) \\
w &= \frac{dL}{dt} \text{ (m/Ma)} \\
T &= 20 + 0.033L \degree\text{C}
\end{align*}
\]
where \( C(x,t) \) is concentration, \( x \) is distance, \( t \) is time, \( L(t) \) is the depth to top of the evaporite, \( \varnothing \) is porosity, \( T \) is the temperature at the top of the evaporite, \( s(T) \) is the solubility of halite at temperature \( T \), \( w \) is the subsidence rate, \( a \) is a constant, and \( D(x,t) \) is the sediment diffusion coefficient with units such that the flux:

\[
J(x,t) = - D(x,t) \frac{\partial C}{\partial x} \text{(mass unit area}^{-1}\text{.time}^{-1})
\]

For simple cases where the diffusion coefficient is constant, the diffusion equation can be solved analytically by separation of variables and Fourier series theory (Carslaw and Jaeger, 1959).

However, when the effects of porosity, temperature, viscosity, and concentration on the diffusion coefficient are taken into account, the diffusion coefficient becomes a highly complicated function of depth and time. Because of this, an analytical solution cannot be obtained. Finite difference methods are useful in such cases and provide approximations to the solution of eqn. (1) at discrete values of depth.

The finite difference equations express approximate derivatives of concentration at discrete depths. We have used a Lagrangian reference frame by fixing the position of the nodes relative to the subsiding sediment. Consequently, we can solve the one-dimensional diffusion equation without having to include an advective term.
In our calculations, we have not taken into account the impact of a non-equilibrium temperature gradient because its impact is likely to be secondary in importance to the effects of sediment porosity. Because the sediment diffusion coefficient is directly proportional to absolute temperature (°K), even a 50 °K difference between equilibrium and nonequilibrium temperatures at a given depth would increase the sediment diffusion coefficient by a small amount. For example, in the specific case when the equilibrium temperature is 350 °K (77°C), the diffusion coefficient would be raised by only 15% if the non-equilibrium temperature was 400 °K.

Initial and Boundary Conditions

We consider the one-dimensional diffusive mass transport of Na and Cl upward from a halite bed being progressively buried by sediments which initially contain fresh connate water. We use "connate" here in the strict, original sense of Lane (1907): water initially buried in the sediment at the time of deposition. When this water is modified in composition or is physically displaced, it is no longer connate. In the case of our actively subsiding sedimentary basin, the NaCl concentration is fixed at the upper boundary at 0 g l⁻¹. At the lower boundary, which coincides with the top the evaporite, the NaCl concentration is determined by halite saturation at the appropriate depth. At 20°C, a saturated solution of NaCl contains 26.5% by weight NaCl and has a
density of 1.20 g cm\(^{-3}\) (Weast and Astle, 1983, p. D-262). This value is also close to the salinity of pore waters in the Michigan Basin at a depth of 3000 m.

In many sedimentary basins dominantly filled with sand, porosity decreases exponentially with depth such that:

\[ \phi = \phi_0 e^{-\alpha x} \]

where \(\phi_0\) = porosity at \(x = 0\) m, and \(\alpha\) = a constant, with units of \(m^{-1}\), (Athy, 1930; Dickinson, 1953; Thomas and Oliver, 1979; Hoholick et al., 1984). The above equation is justified because it very roughly represents an average for a variety of measurements in sands and shales from the Northern Gulf Coast (Fig. 2.3).

**The moving boundary problem - case of a subsiding evaporite**

From Taylor's formula, the backward difference approximation for \(\partial C/\partial t\) at the \(i\)th step in depth and the \(j\)th step in time is

\[ \partial C/\partial t = (C_{i,j} - C_{i,j-1})k^{-1} \]

where \(k\) equals the time increment. Because sediment with fresh connate water is continuously being added at the top, a point at the \(i\)th step in depth at the \(j\)th step in time is actually at the \((i-1)\)th step in depth
Fig. 2.3. Decrease in porosity with depth for Gulf Coast sands and shales (after Hanor, 1979). The solid lines bound porosities of sands while the broken lines bound porosities for shales. The porosity-depth curve $\phi = 0.5e^{-0.0003x}$, where $x$ is in meters, is represented by points. The porosity-depth curve $\phi = 0.8e^{-0.0012x}$ is represented by triangles.
at the $j-1$th step in time (Fig. 2. 4). Therefore, the appropriate backward difference approximation is:

$$\frac{\partial C}{\partial t} = (C_{i,j} - C_{i-1,j-1}).k^{-1}$$

A mass conservative finite difference approximation for the right-hand side of eqn. (1) can be derived as follows

$$\phi_i^{-1}.\frac{\partial}{\partial x}(D(x,t).\frac{\partial C}{\partial x}) = \phi_i^{-1}.h^{-2}.\{D_{i+1/2,j} (C_{i+1,j} - C_{i,j}) - D_{i-1/2,j} (C_{i,j} - C_{i-1,j})\}$$

where:

$$D_{i+1/2,j} = (D_{i,j}.D_{i+1,j})^{0.5}$$

$$D_{i-1/2,j} = (D_{i,j}.D_{i-1,j})^{0.5}$$

and $h$ represents the increment in distance. Substitution of these finite difference approximations into eqn. (1) yields a system of $m-2$ equations with $m-2$ unknowns at any given step in time $j$:

$$A_{i,j}[C_{i+1,j}] + B_{i,j}[C_{i,j}] + E_{i,j}[C_{i-1,j}] = k^{-1}[C_{i-1,j-1}] \quad (3)$$

where:

$$A_{i,j} = - (D_{i,j}.D_{i,j+1})^{0.5}.\phi_i^{-1}.h^{-2}$$

$$B_{i,j} = (D_{i,j}.D_{i+1,j})^{0.5}.\phi_i^{-1}.h^{-2} + (D_{i}.D_{i,j})^{0.5}.\phi_i^{-1}.h^{-2} + k^{-1}$$
\[ E_{ij} = -(D_{ij}-D_{i-1,j})0.5\cdot\rho_{i-1}\cdot h^{-2} \]

for \( i = 2, 3, 4, \ldots, m-1 \); and \( j = 1, 2, 3, \ldots, n \).

The system of equations has a tridiagonal, banded coefficient matrix and can be most efficiently solved by using an algorithm for Crout reduction, commonly referred to as the Thomas algorithm (Burden et al., 1981, p. 312). The method yields solutions that are first order accurate in time, second order accurate in depth, and unconditionally stable. For the special problem of diffusion between non-moving boundaries, for which an analytical solution can be derived, the numerical solution is accurate to 5 or 6 significant figures. For the moving boundary problem, no analytical solutions exist to our knowledge. Because the time step and depth increment between nodes are important factors controlling the accuracy of the numerical solutions, we have tried internodal spacings ranging from 5 to 50 m and found that the solutions did not change significantly, that is, within 3 or 4 significant figures. Therefore, in our calculations, we set the internodal spacing at 10 m. As a result, there were 301 nodes after the evaporite had subsided to a depth of 3000 m.

The resultant solution yields values for transient concentrations of dissolved salts in pore waters. The procedure for the case of an actively subsiding sedimentary basin with a basal evaporite layer can be graphically explained by referring to Fig. 2.4. At time \( t=0 \), the surface concentration \( C_{1,0} \), and the concentration at the top of the evaporite \( C_{4,0} \), at a depth of 3 h (m) are specified, as well as values for \( C_{2,0} \) and \( C_{3,0} \).
Fig. 2.4. Finite difference grid for solving the diffusion equation with a moving lower boundary. The lower boundary which coincides with the top of the evaporite subsides at a constant rate of $h$ meters per k Ma. A point at the $i$th depth at the $j$th time step was actually at the $i-1$th depth at the $j-1$th step in time.
which follow from the initial condition. In a following section, we will show how values for \( D_{i,j} \) are calculated for \( i = 2, 3, 4, \ldots, m-1 \). At \( j = 1, t = k \). \( C_{1,1} \) is known from the upper boundary condition and \( C_{5,1} \) can be calculated from the subsidence rate and the temperature at depth \( x = 4 \) h (m). Substitution of \( i = 2, i = 3, \) and \( i = 4 \) into eqn. (3) yields three equations with three unknowns, \( C_{2,1}, C_{3,1}, \) and \( C_{4,1} \) which are solved simultaneously. The procedure is repeated at \( j = 2 (t = 2k), j = 3 (t = 3k), \ldots, \) up to \( j = n (t = nk) \). At each time increment from \( t = j.k \) to \( t = (j+1).k \), one node is added at the upper boundary. Consequently, there is one additional equation and one additional unknown.

**Flux of salt integrated through time**

Fick's first law states that the flux due to diffusion is:

\[
J(x,t) = -D(x,t) \frac{\partial C}{\partial x}[C(x,t)]
\]

where \( j = \) flux (g.m\(^{-2}\).Ma\(^{-1}\)) and \( \frac{\partial C}{\partial x} \) is the concentration gradient. Because concentrations are now known at \( t = k \), for \( i=2, 3, \ldots, m-1 \), the cumulative flux of NaCl from \( t = 0 \) to \( t = k \) can be calculated using the difference equation:

\[
J_{i,j} = -D_{i,j}[C_{i+1,j} - C_{i-1,j}] \cdot (2h)^{-1}
\]

The procedure for calculating concentrations is repeated when \( j = 2 (t = 2k), j = 3 (t = 3k), \ldots, \) and \( j = n (t = nk) \). At each time step, the cumulative
flux is obtained for elevation levels fixed with respect to the surface, where \( i = 2, 3, \ldots, m-1 \), by adding the mass transferred for that time increment to the total amount of mass transferred for all previous time increments.

As will be demonstrated, a significant amount of salt remains dissolved in the pore water column and this amount increases as the basin subsides. The total amount of salt \( S \) remaining in a pore water column of height \( X \) m and horizontal cross-sectional area \( 1 \) m\(^2\) is

\[
S = \int_{0}^{X} \sigma(x) \cdot C(x) \, dx
\]

or

\[
S = \int_{0}^{X} 0.5(\sigma_{i+1} + \sigma_{i}) \cdot 0.5(C_{i+1} + C_{i}) \, dx
\]

Because \( \sigma(x) \) is known and \( C(x,t) \) can be solved for at each step in time, a numerical integration is performed on the above. The total thickness, \( H(t) \) of compacted halite \((\rho = 2.17 \text{ g cm}^{-3})\) that has dissolved to yield the total amount of salt as NaCl remaining in a pore water column of horizontal cross-sectional area \( 1 \) m\(^2\) at any given time is calculated from the relation:

\[
H(t) = \frac{S}{(2.17 \times 10^6)} \text{ m}
\]
Steady-state concentrations

If the evaporite subsides at a rate of 20 m/Ma and we assume that no further subsidence occurs after 150 Ma, then steady-state concentrations can be calculated for the case when the top of the evaporite is fixed at a depth of 3000 m using the following boundary conditions:

\[ C(0,t) = 0 \text{ g l}^{-1} \text{ NaCl} \]

\[ C(3000,t) = 412 \text{ g l}^{-1} \text{ NaCl} \]

Equation (1) is solved for steady-state conditions, that is, for \( \partial C/\partial t = 0 \). We have assumed that after 150 Ma, the changes in sediment diffusion coefficients with time are small and therefore, the values of D used to generate the tridiagonal coefficient matrix for the system of equations are those for 150 Ma.

Variation of the diffusion coefficient with depth

The sediment diffusion coefficient has a smaller value than the diffusion coefficient in free solution for two reasons: (1) in a horizontal 1 m² cross-sectional area of sediment, the cross-sectional area of pores \( \varnothing \) m² is always less than 1.0; and (2) pores in a sediment are tortuous and not straight tubes. Tortuosity, \( \theta \), is defined as the ratio of the mean length \( L_p \) of the path through the pore space between two points to the straight
line distance between the same points (Lerman, 1979, p. 90). The
dimensional values of tortuosity \( \theta = L_p/L > 1.0 \) (m pore space/ m
sediment); porosity \( 0 < \phi < 1.0 \) (m pore space\(^3\) / (m sediment\(^3\)).

According to Berner (1980, p. 36), the sediment diffusion coefficient \( D \)
used in the previous equations is related to the diffusion coefficient in
free solution \( D^* \) by the relation

\[
D = D^* \cdot \theta^{-2} \quad \text{(length of sediment)}^2/\text{time} \tag{5}
\]

Manheim and Waterman (1974) measured tortuosities for a variety
of sandy and marine sediments and calcareous oozes and found that \( \theta^2 \)
for muds is about twice as great as \( \theta^2 \) for sands, but certainly not orders
of magnitude greater. For some sediments with porosities of 25 - 30%,
500m below the sediment-water interface, \( \theta^2 \) is approximately 3.4. At this
depth, from eqn. (5), \( D = D\theta^2 \), and above 500 m, \( D < D^* \theta^2 \).

According to Lerman (1979, pp. 91-92) the following is an average
for both sands and unconsolidated muds:

\[
D = D^* \theta^2 \quad \text{(length sediment)}^2/\text{time} \tag{6}
\]

and a general equation for shales is:

\[
D = D^* \theta^n \quad \text{(length sediment)}^2/\text{time}
\]
where \( n \) can vary between 2.5 and 5.4. In unconsolidated sands \( n < 2 \), whereas in shales, \( n = 4 \) may be a more appropriate relation. For most of our calculations, we have taken \( n = 2 \), and for a dominantly sandy section this is probably more appropriate. As will be demonstrated later, the choice of \( n = 2 \) is justified because most of the vertical diffusive mass transport past the surface occurs early in the subsidence history, and before the sediments overlying the evaporite are greatly compacted. However, a significant amount of mass transport occurs even in shaly sections and to demonstrate this, we show some sample calculations for \( n = 4 \).

According to Lerman (1979, table 3.4, p. 89) the mean molecular diffusion coefficient for \( \text{NaCl} \) at 25°C in free solution is \( 1.5 \times 10^{-5} \text{ cm}^2 \text{s}^{-1} \). At 20°C, this is equivalent to the relation

\[
D^* = 4.1 \times 10^3 \text{ (m}^2 \text{ Ma}^{-1})
\]

Because \( D^* \) varies with temperature, pressure, and salinity, and because \( \phi \) and \( \theta \) vary with degree of burial compaction, it is obvious that \( D \) will progressively vary with depth and time in a subsiding basin. The following techniques were used to estimate values for \( D^* \) and \( \theta \), and thus \( D \). \( D^* \) for \( \text{NaCl} \) was assumed to vary according to the Stokes - Einstein equation which relates \( D^* \) to changes in temperature and viscosity (Lerman, 1979, p.86). Temperatures were assumed to increase linearly from 20°C at the surface by 33°C km\(^{-1} \), an average geothermal gradient for North Louisiana (Jones, 1975). In-situ viscosities were calculated from
salinity and temperature using the algorithm of Phillips et al. (1981, p.5). The effects of pressure on viscosity and $D^*$ are known to be small, and therefore, pressure was neglected as a variable. Changes in porosity with depth were determined from eqn. (2).

RESULTS

Differential advection of pore water and sediment grains

If steady state compaction of the sediment occurs with burial, that is, if $\partial \rho / \partial t = 0$ for a fixed depth, pore water actually advects downward at a slower velocity than the surrounding solid medium (Berner, 1980, p. 21). This differential advective transport should, in theory, increase the calculated concentrations, concentration gradients, and fluxes that result from diffusion alone. Our calculations (not shown here) indicate that this difference in advective velocity is small early in the subsidence history of the evaporite, and that the advective mass transport which should result from compaction is insignificant in comparison to diffusive mass transport. When the evaporite subsides at a rate of 20 m.Ma$^{-1}$, pore waters are advected downward at velocities of 85-100% of the subsidence rate for times less than 50 Ma. As will be shown, 70-80% of the total thickness of salt dissolved is lost in an early stage of the subsidence history.
Changes in $D$ and viscosity with depth and time

The most important parameters affecting diffusive mass transport past the surface and the salinity of pore waters are: (1) the subsidence rate of the top of the evaporite; and (2) the relation between the sediment diffusion coefficient and the tortuosity, that is, the value of $n$ in $D = D^* \sigma^n$. Figure 2.5 shows how the sediment diffusion coefficient, in units of $10^3$ (m$^2$-sediment).Ma$^{-1}$, changes as a function of depth and time when $D = D^* \sigma^2$. Increases in porosity and temperature increase $D$, whereas increases in viscosity increase $D$. Tortuosity, however, is the main control on $D$. In Fig. 2.5 the maximum value of the diffusion coefficient is about $1.1 \times 10^4$ at a depth of 600 m whereas the minimum value is only $3.6 \times 10^3$ at a depth of 3000 m. At any given depth, the value of $D$ increases with time because, as diffusion of NaCl proceeds, the salinity decreases and a corresponding decrease in viscosity results. At shallow depths, 600 - 1400 m, the effect is almost to double the diffusion coefficients.

Figure 2.6 shows how the viscosity of pore water changes as a function of depth and time. Early in the subsidence history of the basin, at 30 Ma or less, viscosity increases almost continuously with depth to 1.35 c.p. (centipoise) at a depth of 600 m. Although the increase in temperature decreases viscosity, these shallow waters are very saline, and high salinity raises viscosity. As time proceeds, the salinity of pore waters at any given depth decreases partly because of diffusive mass transport upward, and partly because of advection of fresh water downward. Consequently, viscosity initially decreases with depth and
Fig. 2.5. The sediment diffusion coefficient as a function of depth and time for the case when the top of the evaporite subsides at a rate of 20 m/Ma; \( \sigma_o = 0.5 \); \( D = D^* \sigma^2 \); \( C(0,t) = 0 \) g l\(^{-1}\) NaCl; and \( C(3000,t) = 412 \) g l\(^{-1}\) NaCl.
Fig. 2.6. Viscosity of pore water as a function of depth and time. Subsidence rate = 20 m/Ma; $\sigma_0 = 0.05$; $\sigma = 0.0003$; $D = D^* \sigma^2$; $C(0,t) = 0 \text{ g l}^{-1} \text{ NaCl}$; and $C(3000,t) = 412 \text{ g l}^{-1} \text{ NaCl}$.
then increases. For times 50 Ma and greater, the curve becomes C-shaped because of the complex non-linear relation between viscosity and NaCl concentration. Note that as time proceeds, the degree of curvature of the viscosity curve decreases because salinities at shallow and intermediate depths have been considerably lowered as Fig. 2.6 illustrates.

**Changes in salinity with time**

Figure 2.7a shows how the NaCl concentration of pore waters changes as a function of depth and time when $D = D^* \sigma^2$ and the subsidence rate is 20 m.Ma$^{-1}$. For the first 50 Ma or so of subsidence, salinity increases almost linearly with depth, but after 50 Ma, the increase in salinity with depth is curvilinear and the rate of change in salinity with depth increases deeper in the section. This is due to two processes that operate simultaneously: (1) advection of fresh water downward at a rate that is dependent on the rate of subsidence of the evaporite, and (2) thickening of the "diffusion front" above the evaporite. Because sediment diffusion coefficients are low in the deeper part of the section, NaCl cannot be transported upward as fast as at shallow depths, and deep pore waters remain concentrated in NaCl. Figure 2.7a shows that after 150 Ma, NaCl concentrations coincide with steady-state concentrations.

Figure 2.7b illustrates the results when the subsidence rate is 50 m.Ma$^{-1}$. In this case, after only 20 Ma, the salinity profile becomes curvilinear, and after 60 Ma, when the top of the evaporite has subsided
Fig. 2.7. NaCl concentration versus depth. (a) Evaporite subsides slowly at a rate of 20 m/Ma; the salinity profile after 150 Ma coincides with the steady-state profile calculated when the top of the evaporite is fixed at a depth of 3000 m. (b) Evaporites subsides at an intermediate rate of 50 m/Ma; dashed line indicates the steady state profile calculated when the top of the evaporite is fixed at a depth of 3000 m. (c) Evaporite subsides at a depth of 500 m/Ma; dashed line indicates the steady state profile calculated when the top of the evaporite is fixed at a depth of 3000 m, note the large deviation from steady-state when the evaporite has subsided to a depth of 3 km. $\beta_0 = 0.5$, $\alpha = 0.0003$; $D = D \sigma^2$; $C(0,t) = 0 \text{ g l}^{-1} \text{ NaCl}$; and $C(3000,t) = 412 \text{ g l}^{-1} \text{ NaCl}$. 
to a depth of 3000 m, NaCl concentrations have not reached steady-state. Figure 2.7c is analogous to Figure 2.7a and b, but illustrates the results when the subsidence rate is very fast, 500 m.Ma\(^{-1}\). Salinities increase very slowly with depth, at first, and thereafter increase with depth more rapidly. The reason for this is that fresh connate waters are advected downward rapidly. The effect of this is to advect the whole salinity profile downward. When the top of the evaporite is at a depth of 3000 m, NaCl concentrations show a very large deviation from steady state, mainly because rapid subsidence of the evaporite leaves a trail of fresh pore waters above it and there is little time for diffusion to raise NaCl concentrations in the shallower part of the section.

As subsidence becomes more rapid, transient salinity profiles deviate increasingly from the steady-state profiles (Figs. 2.7b and 2.7c). However, for these cases, steady-state salinity profiles are achieved after 150 Ma, implying that diffusive salinity profiles in old evaporite basins may not reflect the early subsidence history. For moderate to high subsidence rates (50 m.Ma\(^{-1}\) to 500 m.Ma\(^{-1}\)) represented by Figures 2.7b and 2.7c, diffusive fluxes of salt past the sediment-water interface would increase after cessation of subsidence, but would still be much less than for the case of a slow subsidence rate, because the NaCl concentration gradient at the sediment-water interface is low. For a given layer, the NaCl concentration increases with time (Fig. 2.7d).
Fig. 2.7 (d) Change in NaCl concentration with time for three specific layers. The subsidence rate is 20 m/Ma and the basin-fill is sand.
Fig. 2.8 Total amount of compacted halite destroyed in meters for a horizontal cross-sectional area of 1 m$^2$ in the sedimentary basin. This has two components: (1) the cumulative flux due to diffusion past a horizontal cross-sectional area of 1 m$^2$ at a depth of 10 m; and (2) the amount of salt that remains dissolved in the pore water, which increases as the evaporite subsides and as the pore water column thickens. The subsidence rate is: (a) 20 m/Ma; (b) 50 m/Ma; and (c) 500 m/Ma. For (a), (b), and (c): $\varphi_0 = 0.05$ (or 5%); $\alpha = 0.003; D = D^* \varphi^2; C(0,t) = 0 \text{ g l}^{-1} \text{ NaCl}$; and $C(3000,t) = 412 \text{ g l}^{-1} \text{ NaCl}$. For (d): $\varphi_0 = 0.8; \alpha = 0.0012; \text{ and } D = D^* \varphi^4$. 
GEOLOGICAL FACTORS CONTROLLING DIFFUSIVE FLUXES

The consequences of diffusive mass transport are important for evaporites in sedimentary basins because an extremely large thickness of salt can be dissolved when the subsidence rate is low. Here, we show calculations of cumulative mass fluxes of NaCl past the sediment water interface made for a variety of different geological situations, using eqn. (4).

Subsidence as a variable

Figure 2.8a shows results for the case when $D = D^*\rho^2$ and when the subsidence rate is 20 m.Ma$^{-1}$. After 150 Ma, 270 m of compacted halite have been destroyed. Diffusive mass transport accounts for 220 m (81%) of the destruction, whereas the amount of salt dissolved and remaining in the pore water column is equivalent to 50 m (18%). The salt flux integrated through time, and the total amount of salt dissolved increased very rapidly during initial subsidence. This is intuitively reasonable because the concentration gradient at a depth of 10 m is high at early times and low in the late subsidence history (Fig. 2.7a, b). Sixty two percent of the salt dissolved is lost in the first 30 Ma, but the rate of salt destruction slows down thereafter because the advection of fresh connate water downward retards diffusion upward to the surface. Late in the subsidence history, upward diffusion of salt merely redistributes salt within the thickening pore water column.
Subsidence histories calculated by Nunn (1984) for Jurassic sediments in the Northern Gulf Coast Basin indicate that between 190 Ma and 160 Ma before the present, the subsidence rate was about 20 m.Ma\(^{-1}\); between 160 Ma and 110 Ma before the present, the subsidence rate averaged 30 m.Ma\(^{-1}\); but for the past 50 Ma, subsidence has been very slow, less than 5 m.Ma\(^{-1}\). Estimates for the post-compaction thickness of the Louann prior to diapirism range from 450 to 600 m (Kupfer et al., 1976). If we take an intermediate thickness of 525 m, and if we assume an average subsidence rate of 20 m.Ma\(^{-1}\) acting for 150 Ma, then 30% of the originally deposited Louann would have to be dissolved and removed by diffusion.

If only a few hundred meters of halite were deposited in a slowly subsiding basin, they would be destroyed during subsidence, as might have been the case in the Illinois Basin, in which bedded evaporites are not observed, but deep pore waters are very salty with as much as 200 g l\(^{-1}\) of total dissolved solids (Fig. 2.1). Additional calculations (not presented) show that if the evaporites are totally destroyed after reaching a depth of 3000 m, and if no further subsidence occurs, the salinity of waters will remain high for a long time, at least 300 Ma after complete dissolution of the evaporites and cessation of subsidence. In evaporites that contain anhydrite, it is possible that significant amounts of Ca may be transported by diffusion as well. The rapidity with which halite can be destroyed by diffusion, given slow subsidence, suggests that K or Mg salts may be destroyed more rapidly because of their higher solubility.
Our calculations do not take into account the fact that, in actual sedimentary basins, there are likely to be sediments under the evaporites, and that diffusion is likely to occur not only upward, but downward as well. We have also neglected the effects of possible overpressuring of sediments below the evaporites at some stage in the subsidence history or possible free convective groundwater flow. If this occurred, a term due to advective flux would have to be added on to the diffusive flux and the net effect would be to increase the amount of halite destroyed. Early in the subsidence history of the North Louisiana Salt Basin, geothermal gradients may have been 60 to 70°C km\(^{-1}\) (Nunn et al., 1984), twice as high as the present day geothermal gradient, 33°C km\(^{-1}\). Intuitively, we should expect diffusive fluxes to be much greater for higher geothermal gradients.

Figure 2.8b shows that the total amount of salt dissolved and the total integrated salt flux past 10 m depth, through time, are lower when the subsidence rate is lower, 50 m Ma\(^{-1}\). This is due to two factors: (1) Fresh connate waters are advected downward at a faster rate, 50 m Ma\(^{-1}\) compared to 20 m Ma\(^{-1}\) in Fig. 2.8a. Therefore, at any given time, concentration gradients near the surface are lower than in Fig. 2.8a. (2) Diffusion acts over a shorter time than in Fig. 2.8a. In this case of more rapid subsidence, the total thickness of salt destroyed after 60 Ma is 112 m of which 71 m (62%) is due to diffusion through the surface and 42 m (38%) is due to salt existing in the pore waters. Figure 2.8c shows the results when the subsidence rate is 500 m Ma\(^{-1}\). The total thickness of salt destroyed is 17 m of which only 14.6 m (86%) is represented by salt.
existing in the pore water column and 2.1 m (14%) is due to diffusion. The small contribution of diffusive mass transport is due to the greatly reduce time factor, 6 Ma in Figure 2.8c versus 60 Ma in Figure 8b and 150 Ma in Figure 2.8a, as well as the rapid advection of fresh connate pore waters downward from the surface.

Such rapid subsidence rates are probably unlikely in the subsidence histories of most intracratonic basins. Subsidence rates in the Michigan Basin were 4 to 80 m Ma⁻¹ in the center and less near the margins (Nunn et al., 1984, p. 299). The results shown in Fig. 2.8c suggest that destruction of evaporites should be great in intracratonic basins, especially at the margins of the basin where subsidence rates should be slower.

*Upper boundary condition as a variable - The burial of sediments with marine water*

So far, we have assumed that sediments with fresh connate water (0 g l⁻¹ NaCl) are deposited. The results obtained by depositing sediments with marine connate water (33 g l⁻¹ NaCl) are not significantly different. The effect of burying the evaporite under sediments with marine connate water is to decrease the total thickness of salt destroyed and the diffusive salt flux by roughly 5-10% due to the reduced concentration gradients.
Lithology as a variable—diffusion in a shaly section

So far, we have taken $\varphi_0 = 0.5$ and $\alpha = 0.0003$ (m$^{-1}$), values which are appropriate for sands and sandstones in the Northern Gulf Coast. For shales, $\varphi_0 = 0.8$ and $\alpha = 0.0012$ (m$^{-1}$) are probably more appropriate (Fig. 2.3). Also, because shales have higher tortuosities, we have selected $D = D^\ast \varphi^4$. The results in Figure 2.8d show that a significant amount of salt is destroyed even when the evaporites are buried by shales. After 150 Ma, a total of 175 m of salt is destroyed of which 165 m (94%) is the diffusive flux and 10 m (6%) is the amount remaining in the pore water column. The reason for this is that the porosity of shale is higher than the porosity of the sandstones in the shallow part of the section, and therefore the term $\varphi^4$ is larger than $\varphi^2$. This is important because most of the diffusive mass transport occurs very early in the subsidence history of the evaporite.

Our calculations thus indicate that the amount of salt destroyed by diffusive mass transport is likely to be significant not only in sand-rich sections, but also in shale-rich sections.

GRAVITATIONAL STABILITY OF PORE WATER

Many workers have suggested that free convection of groundwater occurs in many sedimentary basins (e.g., Donaldson, 1962; Wood and Hewett, 1982; Blanchard and Sharp, 1985). The basic premise is that, when geothermal gradients are sufficiently high, deeper groundwaters
should be less dense than shallow groundwaters and should ascend. So far, we have neglected effects of free convection, and for this reason, our estimates of upward fluxes of NaCl due to diffusion alone must be considered as minimum values.

Using a relation given by Phillips et al. (1981) to calculate the density of NaCl solutions as a function of pressure, temperature, and NaCl concentration, the profile of pore water density versus depth can be traced as a function of subsidence rate and time. Figure 2.9a shows the results when the subsidence rate equals 20 m.Ma⁻¹. Because the subsidence rate is slow, salt diffuses to shallow depths rapidly enough so that densities consistently increase with depth.

Figure 2.9b shows the results when subsidence is more rapid, 50 m.Ma⁻¹. Again, density increases almost continuously with depth except for large times such as 60 Ma when there is a zone from 100 to 600 m that is very slightly unstable. Figure 2.9c shows the results when subsidence is very rapid, 500 m.Ma⁻¹. The main observation to be made here is that only very rapid subsidence gives rise to a gravitationally unstable pore water column. A thick column of fresh water is formed by rapid advection downward from the surface and this fresh water becomes buoyant as it is heated. We conclude that free convection may not be possible in salt basins of the type considered here undergoing active subsidence until the salt beds have subsided to depths of several kilometers.
Fig. 2.9. Density of pore water as a function of depth and time. (a) Subsidence rate equals 20 m/Ma. Although the effect of temperature alone would make fresh water buoyant at depths of 3000 m, the high salinities ensure that the pore water column will be in hydrostatic equilibrium throughout the subsidence history of the basin. (b) Subsidence rate equals 50 m/Ma. Note that there is a zone of slight instability of pore water which extends from 100 to 600 m depth after 60 Ma. (c) Subsidence rate equals 500 m/Ma. Note the zone of gravitationally unstable pore water extends from 100 to 1800 m depth after 6 Ma.
Salinity profiles due to steady-state upward flow plus dispersion

Regardless of subsidence rates, profiles of salinity versus depth due to diffusion have gradients that actually increase downward (Fig. 2.1). The difference implies that although diffusion is an important brine-forming process, by itself, it is inadequate to explain observed salinity profiles in many sedimentary basins. Upward advection of waters with accompanying diffusion and dispersion are likely to be modifying processes. Here, we consider the effects of steady-state upward flow and dispersion on the salinity profile. Bredehoeft and Papadopoulos (1965) noted that upward flow of groundwaters can result in slightly convex upward profiles of temperature versus depth. Here, we make use of their observation to demonstrate the effects of steady-state upward flow and dispersion on the salinity profile.

The effects of upward flow is to shift NaCl concentrations towards high values in the deeper part of the section and thereby simulate salinity profiles actually observed in the Alberta, Illinois, and Michigan Basins (Fig. 2.1). We can evaluate the effects of steady state upward flow by solving the following boundary value problem for NaCl concentrations:

\[ \frac{\partial}{\partial x} \left( D \frac{\partial C}{\partial x} \right) + v \frac{\partial C}{\partial x} = 0 \]

subject to \( C(0) = 0 \text{ g l}^{-1} \) NaCl and \( C(3000) = 412 \text{ g l}^{-1} \) NaCl. This is the advection-dispersion equation for steady-state conditions, where the
advection rate $v$ and the dispersion coefficients $D$ are constants. The solution to this differential equation is

$$C(x) = 412 \cdot \frac{\exp(xv/D) - \exp(3000v/D)) \cdot [\exp(3000v/D) - 1]}{1}$$

and is equivalent to the solution of Bredehoeft and Papadopoulos (1965) for temperatures in a vertical profile. The results are shown in Fig. 2.10 for constant upward flow velocities ranging from -0.01 to -1 m.yr$^{-1}$ and a constant dispersion coefficient of $10^8$ m$^2$.Ma$^{-1}$. Although our choice of a numerically high value of $D$ may be questioned and although constant upward flow velocities may be unrealistic, the solutions clearly demonstrate the importance of upward flow and dispersion in generating convex-upward salinity profiles that are actually observed in some sedimentary basins. Bethke (1985) showed that upward flow velocities due to compaction in intracratonic basins such as the Illinois Basin are on the order of $10^{-3}$ m.yr$^{-1}$. Such velocities may be too low to generate upward-convex salinity profiles shown in Fig. 2.10, but in the Gulf Coast Basins, where sedimentation rates may have been much higher, compaction flow may have been more important.

**SUMMARY AND CONCLUSIONS**

Using finite difference methods, we have shown that, in theory, very saline pore waters can form in subsiding sedimentary basins by dissolution of evaporites and the subsequent upward diffusion of dissolved salt. The mass fluxes are sufficient to destroy several hundred
Fig. 2.10. Salinity profiles due to steady-state upward flow with dispersion. The velocity $v$ is constant for each profile and the dispersion coefficient $D$ is also a constant equal to $10^8$ m/Ma. Results are shown for $v = -0.01$ m/yr, $v = -0.1$ m/yr, $v = -0.5$ m/yr, and $v = -1$ m/yr. The negative velocities take into account the fact that flow is upward. Note the similarity in the upward convexity of these profiles with those for the Alberta, Illinois, and Michigan Basins in Fig. 2.1. The dots are salinities calculated from the steady state solution to the advection-dispersion equation referred to in the text.
meters of compacted NaCl over periods of 150 Ma or less. Where subsidence rates are slow or negligible, a much greater thickness of evaporite could be destroyed, perhaps as much as 1000 m over a 300 Ma period of time. This may be the reason why no evaporites are found in the Illinois Basin although the waters are still very saline at depth. In the North Louisiana Salt Basin, as much as 30% of the salt originally deposited may have been destroyed by the diffusive mass transfer since the time of deposition.

Where subsidence rates are very rapid, fresh waters are advected downward so rapidly that the diffusive fluxes of salt through the surface are negligible. Although diffusion of dissolved NaCl is an important brine-forming process during the subsidence of an evaporite, it does not explain the convex-upward nature of salinity gradients that are actually observed in portions of some sedimentary basins. Advection, convection, and dispersion must be additional transport processes that modify the salinity profiles.

ACKNOWLEDGEMENTS

We thank Dave Evans, Craig Bethke, and Grant Garven for suggesting improvements on various versions of this manuscript. We also thank Mary Lee Eggart for drafting the figures.
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CHAPTER 3

DENSITY-DRIVEN GROUNDWATER FLOW
NEAR SALT DOMES

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In press: Chemical Geology
ABSTRACT

Active halite dissolution occurs around numerous salt domes in Louisiana forming brines with as much as 300 g/l TDS, most of which is Na and Cl. We numerically modeled the possible dynamics of: 1) brine-density flow induced by solute concentration gradients, and 2) thermal convection induced by temperature gradients, around hypothetical salt domes. Calculated maximum flow velocities due to brine density flow range from 1 cm/yr to 1 m/yr even in low permeability sediments, much greater than velocities due to thermal convection. Immense quantities of halite can be dissolved by brine density flow, up to 3 km³/Ma, and formation of brine plumes can occur very rapidly, in some cases requiring less than 1 Ma. Maximum flow velocities increase with bulk permeability and decrease with increasing anisotropy. Geopressured waters bleeding upward into the hydropressured zone may affect brine formation in the hydropressured zone, provided permeability of the geopressured zone is sufficiently high.
1. INTRODUCTION

Numerous salt domes exist in the subsurface of the onshore and offshore Louisiana Gulf Coast, and active halite dissolution occurs along the margins of some of these domes. Minor (1925) reported a systematic increase in salinity of waters near two salt domes, the Edgerly dome in Louisiana and the Goose Creek dome in Texas. Around Welsh dome in Southwest Louisiana, the salinities of waters in Miocene sediments surrounding the dome range from 70 to 200 g/l TDS (Bennett and Hanor, 1987). Chemical analyses show that some pore-waters have as much as 300 g/l TDS near Iowa Dome in Southwest Louisiana. Because sediments around Welsh and Iowa Domes were deposited in fluvial, deltaic, or normal marine environments, connate water that existed in the sediments at the time of their deposition probably had salinities no greater than seawater (35 g/l TDS). Evidently, a large amount of NaCl has been dissolved and transported from distances exceeding 5 km within the past 5 to 25 million years. From calculations of the pore-water density field, Bennett and Hanor (1987) deduced that convective groundwater flow should occur around the Welsh dome, and Hanor (1987) came to a similar conclusion for the Iberia dome in South Louisiana.

Using electric logs of deep oilfield boreholes in south Louisiana, Hanor et al. (1986) found that a wedge of fresh water exists from the surface to depths of about 0.5 km, and that this wedge thins to the south, towards the coast. Furthermore, salinities increase with depth to a
maximum of about 150 g/l between 1 and 4 km and then decrease below that (Fig. 3.1). This salinity profile occurs across the entire 320 km width of south Louisiana. Hanor et al. (1986) attributed the high salinity zone to dissolution of NaCl from numerous salt domes at depths where the brines occur. The interval where the decrease in salinity occurs is dominantly shaly and geopressed (with abnormally high fluid pressures).

It is of obvious interest to determine how quickly plumes of dissolved NaCl can form around salt domes when solute concentration gradients drive groundwater flow, and how quickly halite can be dissolved around shallow salt domes when brine-density flow occurs in sediments flanking the salt domes. Natural density-driven groundwater flow may be a consequence of solute concentration gradients as well as temperature gradients. In order to simulate this type of flow, it is necessary to solve three coupled partial differential equations, one for fluid flow, one for solute transport, and one for heat transport, the three equations being coupled by fluid density. At a recent meeting of HYDROCOIN, an international cooperative project for studying, and comparing groundwater modeling strategies for radioactive waste management, the general conclusion reached was that nonlinear problems such as density-driven flow resulting from salt-brine diffusion pose a great difficulty for many numerical simulation codes (Nicholson and Cole, 1987). It is thus not surprising that from a list of 225 documented computer codes for groundwater flow, solute transport, and heat transport (Van der Heijde and Srinivasan, 1985) only two are
Figure 3.1. Fence diagram showing the salinity distribution of pore waters in the subsurface of south Louisiana, at bottom, and location of cross-sections comprising the fence, at top (after Hanor et al., 1986). Note: a) a wedge of fresh water, <1 g/l, thinning from North to South, b) widespread occurrence of brines with >100 g/l TDS at depths >1 km, and c) decrease in salinities at depths of 3 to 4 km. The widespread occurrence of brines with 100 to 150 g/l TDS corresponds to depths to which many salt domes have risen.
capable of simultaneously solving all three coupled equations mentioned above. The first of these codes (Runchal and Hocking, 1981; Fujioka and Runchal, 1982) is proprietary, whereas the second, SWIFT (Reeves and Cranwell, 1981; Finley and Reeves, 1982), was not available to us in a compatible version at the time we started this project. To our knowledge, the only geologic example in which brine density flow is fully coupled with heat transport, is a study of wastewater injection in a coastal aquifer by Fujioka and Runchal (1982). In the case studied by these workers, the range of salinities and temperatures was very low.

Using SUTRA, a U.S. Geological Survey numerical model for simultaneous fluid and heat or solute transport (Voss, 1984), we have investigated the possible dynamics of density-driven groundwater flow near salt domes. The advantages of SUTRA are that it permits simulation of flow due to gradients in either solute concentration or temperature. In one case, we permit density-driven flow to occur in response to solute concentration gradients arising from halite dissolution around a salt dome, and neglect the effects of temperature. In a second case, we permit density-driven flow to occur in response to temperature gradients and neglect the effects of solute concentration. We show how, under certain conditions, large volumes of halite can be dissolved from the top and flanks of salt domes. Also, we show that the magnitudes of flow resulting from thermal convection are likely to be less than that of brine-density flow by a factor of 0.05 for similar geometries and sediment properties.
2. CONCEPTUAL MODEL OF FLOW AROUND SALT DOMES

Figure 3.2 shows the location of 203 known salt domes and three salt dome basins in the subsurface of onshore and offshore Louisiana. In the North Louisiana Salt Basin, the rate of salt dome growth was most rapid in the late Jurassic, 140 to 145 Ma b.p. (Lobao and Pilger, 1985, p.196) with a period of minor growth in the Cretaceous from 105 to 115 Ma b.p. In south Louisiana, dome growth is currently active, as is evidenced by outcropping of salt or uplift of recent sediments, or was recently active during the Miocene or Pliocene because of very rapid sediment deposition during these times.

In numerous areas of south Louisiana, shallow salt diapirs have penetrated upward through a deep sequence of overpressured marine shales, sediments with relatively low permeability, into an overlying sequence of fluvial and deltaic sands. It is in this uppermost, more permeable section that salt dissolution and brine formation appear to be occurring. In the numerical calculations described here, we have modeled flow conditions around a shallow dome where the top of the salt is 1 km deep and the base of the permeable section is 2 km deep. Within salt dome basins, domes appear to be spaced 10 to 20 km apart (Fig. 3.2). Therefore, for simulation purposes, the dome of interest is assumed to be part of an array of similar domes spaced 10 km apart (Fig. 3.3), and thus has a radius of influence of 5 km. In fact, the cross-sectional area of influence of each dome in the array is a hexagonal polygon. By treating this hexagonal polygon as a circle, we consider flow in only two
Figure 3.2. Locations of the 203 known salt domes in onshore and offshore Louisiana (modified after a map prepared by the Department of Natural Resources, Louisiana Geological Survey). The salt domes are located in three provinces: the North Louisiana Salt Basin, the northwestern extension of the Mississippi Salt Basin within Louisiana, and the Gulf Salt Basin. Within each salt basin, the average spacing between domes is 10 - 20 km, although greater or lesser spacings exist for certain domes.
Figure 3.3. The hypothetical dome of interest for simulation purposes, in the center, is surrounded by six identical domes which are 10 km away. The radially symmetric cross-section of interest is shown shaded. In fact, the cross-sectional area of influence for each salt dome is a hexagonal polygon, but for this study, we assume that this hexagonal polygon can be approximated by a circle of radius 5 km. In doing so, we consider flow in only two dimensions. The solute concentration contours and flow vectors are schematic and do not have assigned numerical values. In general solute concentrations decrease upward.
dimensions. While the above interdomal spacing and radius of influence fall within typical ranges observed in much of the Louisiana Gulf Coast, details of the geologic setting of naturally occurring domes are of course unique (Halbouty, 1979). We wish to emphasize that the idealized cases discussed here are not meant to describe specific field examples or a universal model, but rather to represent some of our initial numerical experiments designed to provide general information on probable dynamics of brine-density flow.

3. TYPES OF SIMULATIONS CONDUCTED

We conducted two categories of simulations. In the first, we studied brine density flow neglecting the effects of temperature, whereas in the second, we studied thermal convection neglecting the effects of solute concentration. Our intent was to evaluate the relative importance of the two end-member types of density-driven flow. For all of our simulations, impermeable boundaries were specified at the sides of the cross-section (Figs. 3.4, 3.5, and 3.6), whereas at the surface, fluid pressure was assigned a constant value of 0 Pa. For two simulations which evaluated the effect of fluids escaping from a geopressed (abnormally pressured) zone, abnormally high fluid pressures were specified within the geopressed sediment (Fig. 3.5).

The boundary conditions stated previously are justified by our conceptual model shown in Fig. 3.3 for pervasive brine formation in much of South Louisiana at depths below 1 km (Fig. 3.1). In our conceptual model for
Figure 3.4. Finite element mesh for brine density flow. Stippled area corresponds to salt dome. At the surface, $P=0.0$ and $C=0.0$. No fluid or solute flux occurs out of the sides and base. Nodes within the salt dome have extremely low porosity and permeability ($\varnothing=10^{-4}$ and $K=10^{-12}$ darcy), and are always saturated with respect to halite ($C=0.3$).
Figure 3.5. Finite element mesh for brine density flow with escape of geopressed waters superimposed. The lower layer of elements, lightly shaded, is the caprock of geopressed zone. Nodes with dots indicate constant pressure of 36 MPa, inbetween hydrostatic and lithostatic pressure for a depth of 2 km. The geopressure caprock, is anisotropic, with $k_{\text{min}}/k_{\text{max}}=10^{-2}$, where $k_{\text{min}}$ is minimum permeability (vertical) and $k_{\text{max}}$ is maximum permeability (horizontal). The solid black element has $k_{\text{min}}=k_{\text{max}}$ and therefore acts as a tap bleeding the geopressed zone. No fluid or solute flux occurs out of the sides. At the top, fluid pressure is 0 Pa and solute concentration is 0.
Figure 3.6. Finite element mesh for thermal convection simulations. Top of basement at 8 km is an impermeable barrier. Heat flux normal to the base equals 50 mW/m² (mW=10⁻³ Watt), or 1.2 Heat Flow Units. No fluid or heat flux occurs out of the sides. At the surface, P=0.0 and T=20°C. Porosity and permeability of Units 1 and 2 are referred to given text (see section 3.2). Unit 3 corresponds to the salt dome.
brine formation, brine plumes form around each of the hypothetical salt domes, indicated by the schematic solute concentration contours in Fig. 3.3. As time progresses, plumes coalesce to form brines over a wide area. At the vertical boundaries between coalescing brine plumes from adjacent domes, no lateral fluid or solute flux would occur and groundwater flow and solute transport would be upward. We realize that our assumption of fresh pore water at initial time is probably unrealistic because as a salt diapir pierces sediment, brines would form concomitant to diapirism. However, because this factor is too complex to take into account here, we chose to assume the existence of an initially fresh pore water column. This yields maximum times for brine formation. In order to determine the sensitivity of model results to different geologic factors, the following types of simulations were conducted.

3.1 Solute Transport or Brine Density Flow

The finite element method used in SUTRA yields transient solutions to the flow field and either solute concentrations (for brine density flow), or temperatures (for thermal convection) at discrete points in space. Restrictions on computation time require that the number of points at which a solution is obtained should not be more than a few hundred. As a consequence, the size of the smallest element in our mesh, within which permeability is a constant, is approximately 100 m x 100 m (Fig. 3.4). Therefore, in modeling natural sequences in which individual sedimentary units range from 10 cm to 50 m thick, individual units must be lumped together.
Most measurements of permeability in the subsurface of South Louisiana have been made from cores recovered from oil wells. The sampling frequently reflects a bias towards sands rather than muds because it is the sands that are more permeable, form petroleum reservoirs, and are therefore of greater interest. Permeability measurements of sands in South Louisiana over a depth range of 0 to 2 km range from $10^{-3}$ darcy to 5 darcy according to the data of Bebout and Gutierrez (1981) and measurements from sediments around West Hackberry dome show that the most common values at depths of 1600 m to 2000 m are $10^{-3}$ to $10^{-2}$ darcy and $10^{-1}$ to 1 darcy (Fig. 4.8 in Chapter 4). Permeabilities of less than $10^{-3}$ darcy usually correspond to muds, silts, and cemented sands. Our choice of permeabilities for the simulations that follow is thus consistent with measurements, although on the low side, because it takes into account the sampling bias towards sands.

For solute transport simulations, all nodes within the salt dome were assigned constant NaCl concentrations of 0.3, or 30 wt%, an approximate value for halite saturation at the depth of interest. In all of the cases here, we assume that even though salt dissolves with time, the shape of the salt dome does not change because new salt is continuously fed from beneath. We assigned a porosity of $10^{-4}$ (essentially zero), expressed as a ratio of pore volume to sediment volume, and a permeability of $10^{-12}$ darcy to the dome to make the surface of the dome analogous to a no-flow boundary and minimize
aqueous fluid flow within the dome. At the surface, solute concentration was assigned a constant value of 0.0, based on the assumption that the influx of meteoric water or advective-diffusive loss of NaCl at the surface would keep surface waters fresh. In fact, Hanor et al. (1986) have shown that a surface zone of fresh water is widespread in south Louisiana, even where salt domes are common in the subsurface. For initial conditions, we assumed that fresh (C = 0) pore-waters existed everywhere and were hydrostatically pressured.

3.1.1 Homogeneous, isotropic, low-k medium

The dome is surrounded by homogeneous, isotropic sediment of permeability $10^{-3}$ darcy, a value which lies in the middle of the range for consolidated sandstones (Freeze and Cherry, 1979, p. 29). Sands between 0 and 2 km deep in south Louisiana are mostly unconsolidated and have permeabilities that range from $10^{-3}$ to 5.0 darcy (Bebout and Gutierrez, 1981). Unfortunately, there is a dearth of measurements on the mudstones that are interbedded with these sands. We have assumed that the permeabilities of mudstones that are intercalated within the sands are probably lower than those in the adjacent sands by a factor exceeding $10^2$ or $10^3$. Consequently, the $10^{-3}$ darcy value may be considered a mean for sands and shales within the depth interval of interest for our hypothetical salt dome.
3.1.2 Homogeneous, isotropic, high-k medium.

Homogeneous, isotropic sediment with a permeability of $10^{-1}$ darcy surrounds the dome. Our intent in showing results for this case is to demonstrate the rate at which brine plumes may form in shallow, permeable sediments.

3.1.3 Homogeneous, anisotropic, low-k sediment with a low $k_{\text{min}}/k_{\text{max}}$ ratio and a high $k_{\text{min}}/k_{\text{max}}$ ratio.

In this case, maximum permeability in the horizontal direction is $10^{-3}$ darcy. Two examples are shown, for which $k_{\text{min}}$, minimum permeabilities in the vertical direction are $10^{-4}$ darcy and $10^{-5}$ darcy respectively.

3.1.4 Geopressured flow from low-k geopressed zone

In the Gulf of Mexico basin, maximum velocities generated by compaction-driven flow due to expulsion of geopressed waters is on the order of $10^{-3}$ m/yr (Blanchard, 1987). Bredehoeft and Hanshaw (1968) calculated that hydraulic conductivities of $10^{-8}$ cm/sec or lower would be needed to create fluid pressures approaching lithostatic for a sedimentation rate of 500 m/Ma. This corresponds to a permeability of $10^{-5}$ darcy. In the Mississippi fan, a deep-sea submarine fan in the Gulf of Mexico, permeabilities of overpressured sediments have been reported
to range from $10^{-3}$ darcy for near-surface sediments to $10^{-6}$ darcy for sediments at 550 m sub-sea bottom depth (Bryant et al., 1985, p.299).

This simulation was designed to test the influence of geopressured waters escaping upward into the brine plume. A constant fluid pressure of 36 MPa was assigned to the base of the hydropressured zone for nodes in sediment at a depth of 2 km (Fig. 3.5). This corresponds to a fluid pressure midway between hydrostatic and lithostatic pressure. Our choice of permeabilities for the geopressure caprock in Figure 5 was guided by the observations of Bredehoeft and Hanshaw (1968), and Bryant et al. (1985). Permeabilities of the geopressure caprock between 1900 m and 2000 m are $k_{\text{max}}=10^{-6}$ darcy, $k_{\text{min}}=10^{-8}$ darcy. Homogeneous isotropic sediment with $k=10^{-3}$ darcy exists above 1900 m. Porosity is 0.25 for sediments above 1900 m depth and 0.10 for geopressed sediment from 1900 m to 2000 m depth. We assumed in this simulation that waters influent into the lower boundary are fresh with $C=0$.

3.1.5 Geopressed flow from high-k geopressed zone

This is similar to case 3.1.4, but the permeabilities of the geopressure caprock between 1900 m and 2000 m are higher: $k_{\text{max}}=10^{-4}$ darcy, $k_{\text{min}}=10^{-6}$ darcy. This case evaluates the importance of permeability of the geopressed zone in controlling fluid outflux.
3.2 Thermal Convection

In this category, we assume an impermeable base at a depth of 8 km corresponding to the basement (Fig. 3.6). We assume a basement heat flux of 50 mW/m² (mW = 10⁻³ Watt) corresponding to 1.2 Heat Flow Units. This is a value in accordance with basement heat flux measurements in the northern Gulf Coast (Smith et al., 1981). Thermal conductivity of the salt was specified as 6.0 W/m°C (Keen, 1983; Jensen, 1983), whereas thermal conductivity of water and solid grains were specified as 0.7 W/m°C and 1.8 W/m°C respectively (Wood and Hewett, 1982, p.1710). Because of its high thermal conductivity, the salt dome acts as a conduit for heat flow. For initial conditions, we used a steady state, conductive temperature field with fluid velocities equal to zero.

3.2.1 Two layer case with high-k lower layer

For this case, a porosity of 0.25 was assigned to sediment above 2 km (Unit 1 in Fig. 3.6), and a porosity of 0.05 was assigned to sediment from 2 to 8 km deep (Unit 2 in Fig. 3.6). Here, we made no distinction between permeabilities of Unit 1 and 2 (Fig. 3.6) and set $k_{\text{max}}=10^{-3}$ darcy and $k_{\text{min}}=10^{-5}$ darcy for all sediment surrounding the dome.
3.2.2 Two layer case with low-k lower layer

Here, we set \( k_{\text{max}}=10^{-3}\) darcy and \( k_{\text{min}}=10^{-5}\) darcy for all sediment above 2 km (Unit 1 in Fig. 3.6), whereas \( k_{\text{max}}=10^{-6}\) darcy, \( k_{\text{min}}=10^{-8}\) darcy for sediment from 2 to 8 km deep (Unit 2 in Figure 3.6). This permits us to compare flow fields resulting from thermal convection alone with those resulting from brine-density flow, for the region shallower than 2 km, because velocities in sediment below 2 km deep are minimized.

4. EQUATIONS OF FLOW AND TRANSPORT

For solute transport, two partial differential equations are solved simultaneously, the fluid mass balance equation (Voss, 1984, p.34) and the solute mass balance equation (Voss, 1984, p.42). A solution is obtained at each time step for fluid pressure, \( P \), and solute concentration, \( C \), at each node in the finite element mesh. For heat transport, the fluid mass balance equation is solved simultaneously with the solid matrix-fluid heat balance equation (Voss, 1984, p.38). Here, a solution is obtained at each time step for pressure, \( P \), and for temperature, \( T \), at each node in the mesh.

For solute transport, a constant value of \( 1.5 \times 10^{-9} \) m\(^2\)/sec was used for the mean molecular diffusion coefficient of NaCl in free solution, \( D_0 \). According to Lerman (1979, p.89), this is the value for 20°C. Also, a constant longitudinal dispersivity of 100 m and lateral dispersivity of 10 m
were used for all solute transport simulations, values which lie in the 0 to 300 m range for dispersivities calibrated from field tests (Anderson, 1979, p.127). Within this category, we discuss results of the following five cases.

5. MODIFICATIONS TO SUTRA

Minor modifications were made in SUTRA to make effective sediment diffusion coefficients dependent on porosity, $\phi$. Using relations given by Lerman (1979, p. 91), we assigned a tortuosity numerically equivalent to $\phi^{-1}$ to sediments with maximum permeability, $k_{\text{max}}$ equal to or greater than $10^{-3}$ darcy, assuming that such sediments represent sands. We assigned a tortuosity numerically equivalent to $\phi^{-2}$ to sediments with permeability of less than $10^{-3}$ darcy, assuming that such sediments represent muds. The resultant sediment diffusion coefficient for sands (Berner, 1980, p.36), corrected for tortuosity is thus

$$D = D_0 \times \phi^2$$

and for shales,

$$D = D_0 \times \phi^3$$

For simulations of thermal convection, a modification of SUTRA was made to make thermal conductivities of elements dependent upon lithology.
6. FLUID PROPERTIES

For our hypothetical cross-section of interest (Fig. 3.3), fluid pressures vary from 0 to approximately 20 MPa (200 bars). The increase in fluid density due to pressure alone can be calculated using the data of Phillips et al. (1981), and is approximately 0.8% for a 1 molal NaCl solution. In comparison, the maximum increase in density due to solute concentration alone is 20%, that which results when fresh water is saturated with NaCl. Because of this, we feel it is reasonable to neglect the pressure-dependence of density. Assuming an average geothermal gradient of 30°C/km, a reasonable value in the vicinity of salt domes in south Louisiana, temperature would increase from 20°C at the surface to 80°C at depth of 2 km. Using the data of Phillips et al. (1981), the variation in fluid density with temperature is approximately 2.3% over this temperature interval.

For density-driven flow resulting from solute concentration gradients, fluid density in SUTRA is dependent on solute concentration only (Voss, 1984, p. 18) and the effects of temperature are neglected. Fluid density at concentration C, \( \rho(C) \), can be calculated from

\[
\rho(C) = \rho_0 + \frac{\partial \rho}{\partial C} \times [C - C_0]
\]

where

\( \rho_0 = \text{density of fluid with reference concentration} \)

\( (10^3 \text{ kg-fluid/m}^3) \)
\[ \frac{\partial \rho}{\partial C} = \text{rate of change in fluid density with solute concentration (700 kg-fluid}^2/m^3/kg\text{-solute)} \]
\[ C = \text{concentration of solute (kg-solute/kg-fluid)} \]
\[ C_0 = \text{reference concentration of solute (=0 in this study)} \]

For heat transport simulations, we totally neglect the effects of solute concentration on fluid density, and permit fluid density to be linearly dependent on temperature according to an equation given by Voss (1984, p.17) such that

\[ \rho(T) = \rho_0 + \frac{\partial \rho}{\partial T} \times [T - T_0] \]

where \( \frac{\partial \rho}{\partial T} = \text{rate of change of fluid density with temperature} \)
\[ \rho_0 = \text{density of fluid at temperature } T_0 \]
\[ T = \text{fluid temperature (°C)} \]
\[ T_0 = \text{temperature of fluid at which base density is specified (20°C)} \]

The viscosity of pure water at 25°C is 0.89 x 10^{-3} kg/m/sec whereas the viscosity of a 4 molal NaCl solution at 25°C is 1.38 x 10^{-3} kg/m/sec (Phillips et al.,1981). Viscosity changes due to solute concentration are small in contrast to viscosity changes that result from temperature changes in the range of 20°C to 80°C. Because we were primarily interested in demonstrating order-of-magnitude effects, such as those due to permeability and anisotropy variations, we consider fluid
viscosity to be a constant for all solute transport simulations, and equal to the viscosity of fresh water at 20°C,

\[ \mu_0 = 1.0 \times 10^{-3} \text{ kg/m/sec} \]

Fluid viscosity is given by an equation of the type, \( \mu = \mu(T) \) (Voss, 1984, p.18).

7. RESULTS

7.1 Brine Density Flow

We first present results for the homogeneous, isotropic, low-k sediment with porosity of 0.25, and a permeability of 10\(^{-3}\) darcy. In this case, brine density flow develops almost instantly because dense brines that form near the flank of the salt dome sink, and fresh, meteoric waters flow in to replace them (Fig. 3.7). After 0.1 Ma, maximum seepage velocities of 13 cm/yr occur adjacent to the dome. As time progresses, brines fill up more of the sediment surrounding the salt dome. Consequently, maximum flow velocities decrease, and after 1.0 Ma, the maximum velocity drops to 10 cm/yr. In nature, where brines already exist around salt domes, the velocities may be lower. The velocities calculated at early times are maximum possible velocities; steady-state velocities would probably be one order of magnitude lower. Three stages in the formation of the brine plume are shown, at 0.1 Ma, 0.4 Ma, and 1.0 Ma (Fig. 3.8), and demonstrate the rate at which brines can form. Velocities
Figure 3.7. Brine density flow is induced by the sinking of dense halite-saturated brines formed near the top and flank of the dome. Fresh meteoric waters recharge the sediments above the dome whereas discharge to the surface occurs away from the dome. The large hollow arrow schematically represents vectors for elements immediately adjacent to the dome. Average linear velocity vectors are scaled such that lengths are proportional to \( \log_{10} \) (velocity). The largest vector is plotted at the lower right. An arrow 1/2 the size of the largest arrow has a magnitude \( 10^{-1} \) that of the largest vector, whereas an arrow 1/3 the size of the largest arrow has a magnitude \( 10^{-2} \) that of the largest vector.
Figure 3.8. Formation of brine plumes can be rapid as shown by the solute concentration field for brine density flow in homogeneous, isotropic low-K medium. Solute concentration contours range from $C=0.05$ to $C=0.25$ with a contour interval of 0.05.
are low primarily because of the low permeability chosen here. Because unconsolidated sands have permeabilities $10^2$ to $10^3$ times higher, initial flow velocities might be much larger, at least until diagenesis reduces permeability.

We next show results for the homogeneous, isotropic, high-k case, when $k = 10^{-1}$ darcy. Brine formation is extremely rapid, and a near-steady-state condition is reached after 0.2 Ma (Fig. 3.9). At very early times, velocities near the dome are very high, 10 to 20 m/yr, but diminish to less than 1 m/yr after 0.2 Ma. After 0.2 Ma, solute concentrations and velocities change very slowly with time. Because a constant solute concentration of 30 wt % NaCl is maintained at the sloping flank of the salt dome, there is a lateral density gradient at all times near the salt dome. Consequently, convection continues even after the basin has nearly filled with brine, although the strength of the convection cell indicated by seepage velocities diminishes by more than one order of magnitude.

We show next the results for two anisotropic examples, in which $k_{\text{min}} = 10^{-4}$ darcy, and with $k_{\text{min}} = 10^{-5}$ darcy respectively, and $k_{\text{max}} = 10^{-3}$ darcy. A comparison of the flow field (Fig. 3.10) with that for the homogeneous isotropic low-k case (Fig. 3.7), shows that as $k_{\text{min}}/k_{\text{max}}$ decreases, recharge of meteoric waters from the surface and discharge of waters at large distances from the salt dome axis are reduced. It is also clear that, as $k_{\text{min}}/k_{\text{max}}$ decreases, the formation of brine plumes is
Figure 3.9. Near-steady-state concentrations are achieved in a geologically short time, in a homogeneous, isotropic, high-k medium.
Figure 3.10. In a homogeneous, anisotropic medium recharge from the surface and discharge to the surface is impeded as $k_{\text{min}}$ decreases. Also, as $k_{\text{min}}$ decreases, maximum velocities decrease.
retarded (Fig. 3.11). Maximum velocities also decrease (Fig. 3.10), although not in proportion to the decrease in anisotropy.

Because upwarping of sediments is sometimes observed near the flanks of salt domes, we tested the effects of upwarping by rotating the permeability tensor such that maximum permeability was inclined at 40° clockwise from the X-axis near the salt dome. We found that the effects of upwarping of sediments near the dome on solute concentration and flow fields are minor compared to the those of permeability and anisotropy, and therefore, we have not included figures for this simulation. Brine plume formation and maximum velocities are not significantly changed, although near the salt dome, velocity vectors tend to be inclined in the direction of $k_{\text{max}}$.

The gross rate of growth of a salt dome is defined as the true rate of salt flow, regardless of the actual movement of the diapir crest (Jackson and Seni, 1984, p. 51). The gross growth rates for several salt domes from salt basins of east Texas, north Louisiana and West Germany vary from 0 to 500 m/Ma (Jackson and Seni, 1984). For a salt dome 0.5 km in radius, a net growth rate of 500 m/Ma would require the flowage of 0.4 km$^3$/Ma. A large volume of halite dissolution can be induced by brine density flow in very short periods of time (Fig. 3.12). For permeabilities of $10^{-1}$ darcy, dissolved salt transport is so rapid that more than 2 km$^3$ of halite can be dissolved in the first million years, and as the ratio $k_{\text{min}}/k_{\text{max}}$ decreases, the total volume of salt in the brine plume decreases (Fig. 3.12). The high rates of salt dissolution are partly a
Figure 3.11. In a homogeneous, anisotropic medium, formation of the brine plume is retarded as $k_{\text{min}}$ decreases.
Figure 3.12. Volume of halite dissolved in pore-water as a function of time.
consequence of our assumption of initially fresh waters, and in nature, where brine plumes already exist around salt domes, salt dissolution rates may be lower. Thus, it is conceivable that dissolution induced by brine density flow may balance the gross rate of diapirism in such a manner that net growth rate is zero and the top of the dome remains at the same elevation with respect to the surface.

7.2 Influence of Escaping Geopressed Waters on Brine Density Flow

We now present results for the case of geopressured flow from a low-k geopressed zone, the top of which lies at a depth of 1900 m. The simulations assume constant fluid pressures of 36 MPa at a depth of 2 km within geopressed sediment surrounding the dome (Fig. 3.5). Waters influent into the geopressed zone are fresh, with C=0. Homogeneous, isotropic sediment with a porosity of 0.25 and a permeability of $10^{-3}$ darcy exists in the hydropressed zone, which extends from the surface to a depth of 1900 m. The geopressed zone between 1900 m and 2000 m has a porosity of 0.10. In the first of two simulations shown, the geopressed zone has low permeability, with $k_{\text{max}}=10^{-6}$ darcy and $k_{\text{min}}=10^{-8}$ darcy.

Velocity and concentration fields in the overlying sandy section are very sensitive to the permeability of the geopressed zone. When the permeability of the geopresured zone is low, effluent fluxes from the geopressed zone are small (Fig. 3.13). Consequently, the influence of escaping geopressed waters is small on the flow field in the
Figure 3.13. When the permeability of the geopressed zone is relatively low, the flow field is similar to that for brine-density flow. When the permeability of the geopressed zone is relatively high, there is net discharge at the surface, and the largest flow velocity occurs at the tap bleeding the geopressed zone.
hydropressed zone (Fig. 3.13) and the brine plume descends along the flank of the dome and spreads radially outward (Fig. 3.14). A second example shows results when $k_{\text{max}}=10^{-4}$ darcy and $k_{\text{min}}=10^{-6}$ darcy in the geopressed zone. In this case, flow of waters is upward along the dome, and there is a net discharge of waters at the surface (Fig. 3.13). The brine plume ascends slightly before spreading radially outwards (Fig. 3.14).

7.3 Thermal Convection

In general, flow velocities resulting from thermal convection are lower than those resulting from brine density flow. For the first of these simulations, $k_{\text{max}}=10^{-3}$ darcy and $k_{\text{min}}=10^{-5}$ darcy for the entire sediment column from 0 to 8 km, whereas porosity of 0.25 from 0 to 2 km deep (Unit 1 in Fig. 3.6) and 0.05 from 2 to 8 km deep (Unit 2 in Fig. 3.6). Within 1 Ma, a steady-state temperature field is developed from the initial conductive temperature field (Fig. 3.15). Because the salt acts as a conduit for heat, a negative temperature anomaly is observed near the root of the salt dome and a positive anomaly is observed near the top of the dome. Two convection cells result, a relatively weak cell between 0 and 2 km, and a relatively strong cell between 2 and 8 km (Fig. 3.16). The largest velocities, 9 mm/yr in this case, occur near the basement.

For the second thermal convection simulation, porosity of 0.25, $k_{\text{max}}=10^{-3}$ darcy and $k_{\text{min}}=10^{-5}$ darcy for sediments from 0 to 2 km deep (Unit 1 in Fig. 3.6). From 2 to 8 km deep, porosity is 0.05, $k_{\text{max}}=10^{-6}$ darcy.
Figure 3.14. For geopressed flow from a low-k geopressed zone, halite-saturated brines are formed at the top and edge of the dome, sink and override the top of the geopressed zone. For geopressed flow from a relatively high-k geopressed zone, halite-saturated brines are formed near the tap bleeding the geopressed zone, immediately adjacent to the dome. The brine plume ascends slightly before spreading laterally.
Figure 3.15. Steady state temperature field, reached after 1 Ma, for thermal convection with contours in °C. The salt dome acts as a conduit for heat. Consequently, a negative temperature anomaly of 30°C exists near the root of the dome whereas a positive anomaly of about 8°C exists near the top of the dome. For Unit 1 (refer to Fig. 3.6), \( \alpha = 0.25 \) and for Unit 2, \( \alpha = 0.10 \). For both Unit 1 and Unit 2, \( k_{\text{max}} = 10^{-3} \) darcy and \( k_{\text{min}} = 10^{-5} \) darcy. Boundary conditions are: Top - \( P = 0 \) Pa, \( T = 20°C \); Sides - no lateral fluid or solute flux; Bottom - No fluid flux, heat flux = 50 mW/m². The kink in the 260°C contour is believed to be due to the interpolation scheme used in contouring temperatures.
Figure 3.16. Steady state velocity field, reached after 1 Ma, for thermal convection simulation. For Unit 1 (refer to Fig. 6), $\varphi=0.25$ and for Unit 2, $\varphi=0.10$. For both Unit 1 and Unit 2, $k_{\text{max}}=10^{-3}$ darcy and $k_{\text{min}}=10^{-5}$ darcy.
and $k_{\text{min}}=10^{-8}$ darcy (Unit 2 in Fig. 3.6). We chose to show the velocity field only for the section above 2 km deep because velocities below 2 km deep are very low (Fig. 3.17). Making the section relatively impermeable below 2 km depth allows us to compare the thermal convection velocity field between 0 and 2 km with that for brine density flow for the homogeneous, anisotropic low permeability case (Fig. 3.11). Maximum flow velocities are 0.5 mm/yr, and the sense of motion is opposite to that for brine density flow (Fig. 3.17).

8. PERTINENCE OF RESULTS TO GEOLOGIC OBSERVATIONS

8.1 Brine Formation in the Subsurface of south Louisiana

Results of our numerical modeling show that brine formation can occur rapidly, certainly within the 5 to 25 Ma b.p. time constraint imposed by ages of the sedimentary section containing these brines in south Louisiana. In our models of brine density flow, flow is downward along the flank of the dome and upward away from the dome. In our models of thermal convection, the direction of motion is opposite to that for brine density flow, with flow being upward along the flank of the dome and downward away from the dome. On the flanks of Iberia dome in south Louisiana, Workman and Hanor (1985) postulated that relatively dilute, geopressured waters are bleeding upward along the flank of the dome and dissolving salt mainly off the top. Thus, the resulting sense of motion appears to be similar to that of thermal convection. Complicating the situation at Iberia is the indication that the effect of temperature on the
Figure 3.17. Steady state velocity field, reached after 1 Ma, for thermal convection simulation. For both Unit 1 and Unit 2 (refer to Fig. 6), $\phi=0.25$. For Unit 1, $k_{\text{max}}=10^{-3}$ darcy and $k_{\text{min}}=10^{-5}$ darcy. For Unit 2, $k_{\text{max}}=10^{-6}$ darcy and $k_{\text{min}}=10^{-8}$ darcy. The velocity field is shown only for the upper 2 km of the section because velocities are extremely low below 2 km. The velocity field is shown for only the upper 2 km of the section because velocities are extremely low below 2 km.
density gradient is roughly equal in magnitude to that of solute concentration. One feature which needs to be accounted for in future work is the downward decrease in TDS of south Louisiana brines that Hanor et al. (1986) observed in the geopressed interval, below 3 or 4 km in Fig. 3.1).

Although preliminary calculations indicate that thermal effects are of secondary importance to solute concentration gradients near salt domes where active halite dissolution occurs, we do not wish to dismiss the importance of thermal convection near all salt domes. Whereas brines with more than 300 g/l occur near the Iowa dome in Louisiana, pore-waters around many salt domes in the east Texas Salt Basin rarely have salinities exceeding 10 g/l (Fogg and Kreitler, 1982, p.61). It is conceivable that thermal convection may be of greater importance than brine density flow around salt domes armored against halite dissolution by protective caprock or shale sheaths, or during the early stages of diapir emplacement, before brines have formed (see for example, Rabinowicz et al., 1985).

8.2 Deep Infiltration of Meteoric Waters

Workman and Hanor (1985, p.299) inferred that meteoric waters were descending to depths of 2 km on the flanks of Iberia dome in south Louisiana and mixing with geopressed waters moving upwards along the flanks of Iberia dome. McBride et al. (1987) observed that quartz, anhydrite, and dolomite, and early cements in the Norphlet Formation, a
sand which overlies the Louann Salt, are extremely depleted in oxygen-18 in two wells in the Mississippi Salt Basin. They postulated that deep infiltration of meteoric waters dominated the first 30 to 40 Ma of diagenesis in the Norphlet Sand. Dutton (1986) recorded early cements depleted in oxygen-18 in the Travis Peak Formation of east Texas, suggesting that penetration of meteoric waters to a depth of 2 to 3 km occurred in the early history of rift basins in which these rocks lie. The velocity field that results for the case in which homogeneous, isotropic sediment of permeability $10^{-3}$ darcy surrounds our hypothetical salt dome (Fig. 3.7) suggests that brine-density flow may have had a role in this process. If flow velocities average $10^{-2}$ m/yr along the flank of a dome, it would take $10^5$ years for water to flow from the top of the dome to the base of the section at a depth of 2 km for the examples shown in Figures 3.7 to 3.11.

8.3 Cementation of Sandstones

Wood and Hewett (1982, 1984) showed that mass transport of cementing agents may occur within sandstones by thermal convection. Under certain conditions, this may be an important process near salt domes due to the generation of convective flow by locally high heat flow at the top of the domes. Thermal convection may be especially important where active dissolution of halite is prevented by a caprock or shale sheath, but where active halite dissolution occurs, the magnitudes of brine density flow are much greater than those of thermal convection by a factor of $10^1$ to $10^2$. 
Stuart (1970, p.62) reported the occurrence of halite cement in some sands near Weeks Island salt dome. From a combination of resistivity and porosity logs, a general observation made at two salt domes in Cameron Parish, south Louisiana, is that pervasiveness of cementation increases towards the domes, albeit in limited stratigraphic intervals, and at the base of certain sands (Stuart, 1970, p.62; McManus and Hanor, in press). The origin of the pervasive cements that occur at the base of certain sands near these two salt domes has not been conclusively identified. However, in-situ brine density flow may have played a role in their formation, as Stuart (1970) postulated.

9. SUMMARY AND CONCLUSIONS

Using SUTRA, a U.S. Geological Survey numerical model for variable density groundwater flow, we have theoretically shown the importance of brine density flow around salt domes where active halite dissolution is occurring. Maximum flow velocities of 10 cm/yr occur near the salt dome in homogeneous, isotropic sediments with moderate permeability \((k=10^{-3} \text{ darcy})\). Maximum flow velocities may be greater for uncemented sediments at depths of 0 to 2 km in south Louisiana which have permeabilities as great as 5 darcies. Geopressed waters escaping upward may modify the brine formation in the hydropressed zone if the permeability of the geopressed zone is sufficiently high.
Brine density flow may induce dissolution of up to 2 km³ of halite in 1 Ma, and it may also help explain why meteoric waters sometimes infiltrate deep into sediments flanking the salt domes. In anisotropic sediments where maximum permeability, $k_{max}$, is horizontal and minimum permeability, $k_{min}$, is vertical, decreasing the ratio $k_{min}/k_{max}$ retards brine plume formation.

Our preliminary calculations show that velocities for brine density flow are at least 20 times those for thermal convection. However, for domes armoured against halite dissolution by protective caprock or shale sheaths, thermal convection may be of importance, especially if permeabilities of surrounding sediments are high. Because of its capability for kilometer-scale mass transport in relatively short geologic time, $10^{-1}$ to $10^{0}$ Ma, brine density flow may be of importance in sediment diagenesis.

ACKNOWLEDGEMENTS

We wish to thank Dave Evans for informal discussions concerning solute-driven convection in general, as well as for the use of SUTPLOT, a program used to generate graphic output of SUTRA results. We thank Grant Garven, Harry Posey, and Jim Wood for suggesting improvements to an earlier version of this paper. In addition, we express our thanks to Jeffrey Nunn and Rex Pilger for use of the LSU Geophysics VAX 8200 minicomputer.
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CHAPTER 4

PERCHED BRINE PLUMES ABOVE SALT DOMES
AND DEWATERING OF GEOPRESSURED
SEDIMENTS

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ABSTRACT

Previous studies have shown that there is a brine plume with more than 160 g/L total dissolved solids (TDS) in the subsurface, perched above Welsh salt dome in South Louisiana. The occurrence of this plume has been postulated to be due to the expulsion of geopressed fluids up a fault on the flank of Welsh Dome. The process that resulted in formation of the anomalous brine plume near Welsh Dome is of importance because hydrocarbon production is approximately coincident with the brine plume in map plan, although in cross-sections, the brine plume actually occurs a few hundred meters above hydrocarbon production zones.

Using an existing variable density groundwater flow model, SUTRA, we tested the hypothesis that a pulse of fluid flow up the fault could result in brine formation near the top of the dome. When the permeability of the geopressed zone was very low in our simulations, brine formation in the hydropressed zone occurred entirely by brine-density flow and flow was downward along the fault flanking the dome. Above a certain threshold permeability for the geopressed zone, between $10^{-7}$ darcy and $10^{-5}$ darcy in the vertical direction, geopressed fluids were forced up the fault and brine formation occurred above the dome rather than on the flank. Brines can theoretically form above salt domes in very short geologic times ($10^{-1}$ Ma) by expulsion of waters from the geopressed zone below, but that the fluid effluxes required are
gigantic and appear to be unsustainable over large time intervals such as 10 Ma.

1. INTRODUCTION

The subsurface fluid regime in the salt dome province of South Louisiana is a hydrodynamic one in which immense volumes of halite have been dissolved and NaCl transported over scales of several kilometers. Brines with 100 to 200 g/L total dissolved solids (TDS), most of which is dissolved NaCl, are widespread at depths of 1 to 4 km in South Louisiana (Hanor et al., 1986) and chemical analyses show that waters near some salt domes have more than 300 g/l TDS. Distributions of dissolved salt and volatile fatty acids, trace constituents, in pore waters around salt domes indicate a dynamic subsurface fluid regime (Workman and Hanor, 1985; Hanor, 1987).

Brines with 70 to more than 160 g/L TDS occur in sediments surrounding the Welsh Dome in South Louisiana (Fig. 4.1; Bennett and Hanor, 1987). The sediments which contain these brines were deposited in fluvial or non-hypersaline marine environments and therefore, probably had connate waters with salinities of 0 to 35 g/L TDS (Bennett and Hanor, 1987). More than 6 km$^3$ of halite have had to been dissolved since Miocene time to form the brines around the Welsh Dome (Bennett and Hanor, 1987).
Fig. 4.1 Map of Louisiana showing location of Welsh Dome (after Bennett and Hanor, 1987)
Of special interest here is the occurrence of the most saline brines with more than 160 g/L TDS above and slightly to the north of Welsh Dome (Fig. 4.2) and just above a fault (Fig. 4.4). The origin of this plume has been postulated to be due to the focusing of geopressured fluids in the immediate vicinity of the dome upward and towards the salt dome by a fault (Bennett and Hanor, 1987). In map plan, this plume extends several kilometers laterally away from the dome to the northwest and is approximately coincident with the main area of hydrocarbon production at Welsh (Bennett and Hanor, 1987, p. 659). It is of practical importance to understand the process by which this plume may have formed. Questions we specifically address in this paper are: 1) could a sudden pulse of geopressured fluids escaping up the fault have formed the brine plume perched above Welsh dome, and if so, 2) what are the magnitudes of fluid effluxes required from the geopressured zone to form the brine plume?

Elsewhere in the Gulf Coast, evidence for upward flow along fault planes has been found in South Texas, based on mapping of hydraulic heads (Berg and Habeck, 1982). On a regional scale, updip flow of fluids has been invoked to account for higher than average geothermal gradients in growth fault zones of South Texas (Bodner et al., 1985; Bodner and Sharp, 1988). It is possible of course, that the higher geothermal gradients may also reflect the presence of more thermally conductive sediments, particularly if salt pillows occur at depth.
Fig. 4.2 Salinity distribution around Welsh Dome (after Bennett and Hanor, 1987). Bennett and Hanor (1987) postulated that geopressured fluids escaping up the fault on the north side of the dome may have been responsible for the anomalous plume with more than 160 g/L TDS.
Fig. 4.3 Map of Welsh study area (after Bennett and Hanor, 1987). Solid dots show the location of boreholes used as control. Contours show the depth from ground elevation, which averages 10 m above sea level, to the top of a Lower Miocene sand referred to as the "V-sand". A-A' marks the position of the cross-section in Fig. 4.2.
Fig. 4.4 North - South geologic cross-section of Welsh study area (after Bennett and Hanor, 1987). The lower extent of the fault on the north side of the cross-section is not well known for lack of deep oil wells immediately north of Welsh dome.
Geopressured sediments have pore fluid pressures that exceed hydrostatic fluid pressures and are common in many sedimentary basins (Fertl et al., 1976). Rapid sedimentation and burial of low-permeability sediments is believed to be the primary cause of geopressuring (Dickinson, 1953; Magara, 1971, 1975; Sharp and Domenico, 1976; Sharp, 1978; Keith and Rimstidt, 1985; Bethke, 1986). Smectite dehydration and aquathermal pressuring have also been proposed as a potential cause of geopressuring (Barker, 1972; Powers, 1967; Burst, 1969; Plumley, 1980).

A question of great importance is whether dewatering of geopressured sediments in the U.S. Gulf Coast has occurred slowly and over large time intervals, or whether it occurred episodically as suggested by Cathles and Smith (1983). Cathles and Smith (1983) argued that episodic dewatering of basins with large volumes of muddy sediments was probably responsible for the formation of Mississippi Valley Type (MVT) lead-zinc deposits. They theoretically showed that the high temperatures of formation of these deposits would require upward advection of hot pore fluids through basal aquifers from great depths in a sedimentary basin. The rates of fluid flow required to cause thermal anomalies associated with MVT lead-zinc deposits are more than $10^3$ times greater than could be produced by steady subsidence, sedimentation, or compaction of most basins. Consequently, they suggested that episodic dewatering must occur over time periods representing a few thousandths of the history of basin subsidence. Alternatively, dewatering of geopressured sediments would have had to
occur slowly over large time intervals such as the Eocene to Pliocene, a
time span of approximately 50 Ma, during which thick sequences of low-
permeability sediments were deposited. In this case, the rates of fluid
efflux would have to have been much lower.

In this paper, we use SUTRA, a variable density groundwater flow
model to show that upward escape of fluids through a fault from the
geopressured zone can form brine plumes perched near the top of salt
domes resembling Welsh dome in very short geologic time intervals, $10^{-1}$
Ma, much like the brine plume perched above Welsh dome and bounded
by the 160 g/L TDS contour (Fig. 4.2). We show that molecular diffusion
has probably played an insignificant role in brine formation at Welsh
dome. We evaluate the sensitivity of our model results to variations in the
permeability of the fault and geopressed zone and show that the fluid
fluxes required to form a perched brine plume are extremely large even
when permeability of the geopressed zone is very low.

2. Geology of Welsh Study Area

Welsh dome is the site of one of the oldest oilfields in Louisiana,
having being discovered in 1902 (Reed, 1927). The salt dome is roughly
1.6 km in diameter at its apex, lies at a depth of 2 km, and has no
overhang. There are two main faults in the Welsh study area, an east-
west trending normal fault with 180 m displacement and several smaller
faults with less displacement in the northwest - southeast direction (Fig.
4.3; Bennett and Hanor, 1987, p. 658). A second deep-seated dome,
Woodlawn dome lies 6 km northwest of Welsh at a depth of 3.5 km, but its intrusion has apparently had little or no effect on the structural geology or salinity regime at Welsh (Bennett and Hanor, 1987, p. 659).

As discussed by Bennett and Hanor (1987, p. 656-657), the age of sediments penetrated by oil wells in the Welsh oil field is Oligocene to Recent (Fig. 4.4). Dips of beds below 1200 m depth approach 17° near the margin of the dome but shallower sediments dip more gently 1° or less. The uppermost 300 m of sediment at Welsh consist of Recent sands and muds and coarse Pleistocene sands and gravels. These sediments unconformably overlie a 250 m-thick sequence of Pliocene shales and sands which are underlain in turn by nearly 2 km of Miocene sands, silts, and shales. Below the Miocene section is a 400 m-thick sequence of shales of the upper Oligocene Anahuac (Fig. 4.4). At least 1 km of alternating shales and sands of the Upper Oligocene Frio underlie the Anahuac.

Although the youngest sediments structurally disturbed by faults over the crest of Welsh Dome are mid-Miocene, a topographic mound occurs over the crest of the dome (Reed, 1927), indicating that dome growth appears to be active or that there has been preferential subsidence (Bennett and Hanor, 1987). It is beyond the scope of this study to unravel the history of growth of the Welsh Dome, but we recognize its potential importance in affecting the history of brine formation.
Fluid pressures for wells around Welsh Dome have been calculated from shale resistivities by Bennett (1985, p. 147) and are essentially hydrostatic from the surface to a depth of 3.2 km with a gradient of 10 Pa/m and in excess of hydrostatic below 3.2 km with a gradient of 50 Pa/m (Fig. 4.5). Thin, slightly geopressured intervals occur at depths of 2.2 km and 2.7 km (Fig. 4.5), but these appear to be lenticular zones and not laterally extensive as indicated by their absence in adjacent wells. For this reason, we have not taken into account the effect of thin geopressured intervals within the dominantly hydropressured zone. The top of the main geopressed zone at Welsh corresponds to the top of a thick shale-rich section with less than 10% sand (Bennett and Hanor, 1987).

Much of the Miocene and Anahuac section around Welsh has high salinities, from 70 to more than 130 g/L TDS as far as 10 km from the dome. In the Frio, which is geopressed, salinities decrease with depth from 130 g/L TDS at a depth of 3 km to 70 g/L TDS at a depth of 4 km. In view of low calculated permeabilities of geopressed sediments in the Gulf Coast (Bredehoeft and Hanshaw, 1968; Magara, 1971) it is not clear why salinities of pore waters in geopressed sediments at Welsh, which range from 70 to 100 g/L TDS, are 50 to 80 g/L higher than in chemically analyzed pore waters of geopressed shales at Manchester Field, only 40 km west of Welsh and 16 km southwest of Iowa salt dome (Schmidt, 1973, p. 327). Osmaston (1975) suggested that Schmidt's (1973) analyses for pore waters of geopressed sands represent minimum estimates because Schmidt's (1973) analyses were made by leaching
Fig. 4.5 Vertical profile of fluid pressure versus depth for the deepest well north of Welsh Dome, 4.5 km north of the axis of Welsh Dome. The pressures were calculated from shale resistivities by Bennett (1985, p. 150). The top of the geopressed zone, at a depth of 3.2 km is marked by a sudden increase in the fluid pressure gradient. Within the geopressed zone, the gradient in excess pressure (fluid pressure in excess of hydrostatic) is almost 50 MPa/km, equivalent to a hydraulic head gradient of 5.
shale samples with water and did not take into account the possible precipitation of minerals after recovery of the samples. However, Osmaston (1975) did not elaborate on how better estimates could be obtained.

3. Methods

We assume that the principal driving forces for groundwater flow around salt domes where active halite dissolution occurs are solute-induced gradients in fluid density and gradients in pressure. Although in nature, fluid density variations result from temperature and solute concentration, it appears that solute effects are much larger than temperature effects at Welsh dome. The finite element method used in SUTRA yields transient solutions to the flow field and solute (NaCl) concentrations which are coupled. In more refined simulations, one would have to take into account the coupling of solute transport and heat transport.

In order to simulate the formation of a brine plume under the influence of geopressed flow and solute-induced fluid density gradients, two partial differential equations must be solved simultaneously, the fluid mass balance equation (Voss, 1984, p. 34) and the solute mass balance equation (Voss, 1984, p. 42). A solution is obtained for each time step for fluid pressure $P$, and solute concentration $C$, at each node in a finite element mesh (Fig. 4.6). In these simulations, the temperature field is held constant for reasons which are discussed
Fig. 4.6 Radially symmetric mesh corresponding to the north half of Figures 4.2 and 4.4. Flow units were defined on the basis of the percentage of sand mapped on the north side of Welsh Dome by Bennett and Hanor (1987). Sand percent contours are superimposed. Porosities and permeabilities of flow units are in Table 1. For simulations with geopressure, nodes at the base indicated by dots had $P=75$ MPa and $C=0.02$ (2 wt% NaCl).
### Table 4.1

<table>
<thead>
<tr>
<th>UNIT</th>
<th>LITHOLOGY</th>
<th>PERMEABILITY (darcy) (horizontal)</th>
<th>PERMEABILITY (darcy) (vertical)</th>
<th>POROSITY</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>&gt;50% sand</td>
<td>$10^0$</td>
<td>$10^{-1}$</td>
<td>0.30</td>
</tr>
<tr>
<td>2</td>
<td>25 to 50% sand</td>
<td>$10^{-2}$</td>
<td>$10^{-3}$</td>
<td>0.25</td>
</tr>
<tr>
<td>3</td>
<td>10 to 25% sand</td>
<td>$10^{-3}$</td>
<td>$10^{-5}$</td>
<td>0.25</td>
</tr>
<tr>
<td></td>
<td>(shallow)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>10 to 25% sand</td>
<td>$10^{-3}$</td>
<td>$10^{-5}$</td>
<td>0.20</td>
</tr>
<tr>
<td></td>
<td>(deep)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>&lt;10% sand</td>
<td>$10^{-5}$</td>
<td>$10^{-7}$</td>
<td>0.20</td>
</tr>
<tr>
<td>6</td>
<td>Fault</td>
<td>$10^{-2}$</td>
<td>$10^{-2}$</td>
<td>0.25</td>
</tr>
<tr>
<td>7</td>
<td>Salt Dome</td>
<td>$10^{-12}$</td>
<td>$10^{-12}$</td>
<td>$10^{-4}$</td>
</tr>
</tbody>
</table>

Table 4.1. Seven flow units were defined in the mesh based on lithology. The pattern shown for each corresponds to that used in Fig. 4.6. The lowermost unit with less than 10% sand is geopressed (Fig. 4.6). For some simulations, a horizontal permeability of $10^{-3}$ darcy and a vertical permeability of $10^{-5}$ darcy was used for the geopressed sediment (Unit 5), higher than the values shown above. Also, for some simulations, a higher permeability, $10^{-1}$ darcy was used for Unit 6 which contains the fault.
later. For previous hydrogeological applications using SUTRA, we refer the reader to Souza and Voss (1987), Voss and Souza (1988), and Ranganathan and Hanor (1988).

Previous numerical models of Gulf Coast, basin-scale groundwater flow induced by compaction have concluded that maximum fluid flow velocities due to compaction alone are very small, on the order of $10^{-3}$ to $10^{-2}$ m/yr (Blanchard, 1987; Bethke et al., 1988). Our study differs in scale from those of Blanchard (1987) and Bethke et al. (1988) in three ways. First, the dimensional scale of our field area is almost two orders of magnitude smaller than that of Blanchard (1987) and Bethke et al. (1988). Second, rather than allowing geopressed fluids to migrate large distances laterally (100 km or more), we allow them to be focused upward through a relatively narrow zone 100 m wide with enhanced permeability, within which lies a fault. Finally, whereas Blanchard (1987) and Bethke et al. (1988) conducted their simulations for large geologic time intervals, our simulations are for relatively short geologic time intervals ($10^{-1}$ Ma).

Numerical modeling of variable density groundwater flow and solute transport near salt domes over periods of geologic time more than 1 Ma is difficult for two reasons. First, numerical instabilities and numerical dispersion result if time steps are not small. In using SUTRA, this means that when internodal spacings are 100 m, transport of an imaginary solute particle must not be more than 20 m or so within one time step. As a result, if the maximum velocity near the front of a solute
plume is 1 m/yr, the maximum permissible time step is 20 yrs. This constraint prevents us from conducting simulations for time intervals such as $10^2$ Ma.

4. Conceptual Model of Groundwater Flow near a 'Welsh-type' salt dome

In numerous areas of South Louisiana, shallow salt diapirs have penetrated upward through a deep sequence of overpressured marine shales, sediments with relatively low permeability, into an overlying sequence of fluvial and deltaic sands. It is in this uppermost, more permeable section that salt dissolution and brine formation appear to be occurring. Within salt dome basins of the Louisiana Gulf Coast, domes appear to be spaced 10 to 20 km apart on the average, although greater or lesser spacings do occur. In the numerical calculations described here, we have modeled flow conditions around a 'Welsh-type' dome assuming radial symmetry (Fig. 4.6), and therefore do not consider flow to be three-dimensional. The term 'Welsh-type' refers to the fact that we have used the fault trace on the North flank of Welsh dome (Fig. 4.4), the distribution of sand percentage on the North flank of Welsh dome mapped by Bennett and Hanor (1987) to define the permeability structure, and known fluid pressures in geopressed sediments (Bennett, 1985) as lower boundary conditions.

The underlying assumption for radial symmetry is that, for simulation purposes, the dome of interest is one of an array of similar domes spaced 20 km apart and that each salt dome has an area of
influence bounded by a no-flow vertical barrier at a radius of 10 km. There is some field evidence which may be construed as supporting our assumption of radial transport, at least as a first approximation. Salinity maps around Welsh dome at depths of 1060 m and 2280 m show that there is significant dissolved salt transport in all directions from the axis of Welsh dome outward (Fig. 4.7; Bennett, 1985, p. 36 - 42). In fact, at a depth of 2280 m, the 120 g/l salinity contour covers an area that extends 8 to 10 km radially outward in all directions. There are, however, three-dimensional aspects that we are unable to account for in our simulations, such as the 150 g/l salinity contour at a depth of 1060 m in Fig. 4.7.

In order to account for the three dimensional variations in sediment lithologies and fault geometries, it would be necessary to use a three-dimensional groundwater flow model and solute transport model. The practical difficulties in three-dimensional modeling are: 1) the three-dimensional geometry of sediment bodies must first be mapped and transposed onto a three-dimensional grid which would have to have at least $2 \times 10^4$ nodes (25 nodes or more along the X-, Y-, and Z-axis), and 2) the computations would take very long, possibly $10^2$ times as long as two-dimensional simulations because computation time is proportional to the square of the number of nodes.
Fig. 4.7 Map plan of salinity contours at depths of 1060 m and 2280 m (after Bennett and Hanor, 1987). Note the elongation to the plume bounded by the 150 g/l salinity contour at a depth of 1060 m. Also note that significant amount of dissolved salt transport has occurred in all directions outward from the axis of the dome as indicated by the 120 g/l contour at a depth of 2280 m.
5. Fluid Properties

SUTRA does not permit us to couple heat transport with transport of dissolved salt. Therefore, we have neglected the dependence of fluid density on temperature. In the Welsh study area, salinity gradients are sufficiently large so as to be a stronger control on fluid density than temperature. Salinity ranges approximately from 0 to 160 g/L TDS corresponding to a density range of 1.0 to 1.12. In the Welsh study area, temperature ranges from 20°C at the surface to roughly 120°C at 4 km depth. The corresponding change in density with depth due to temperature alone is about 1/4 the density change due to dissolved salt concentration.

The data of Phillips et al. (1981) show that the effect of pressure on fluid density is very small in comparison with the effect of solute concentration. In the simulations presented here, groundwater flow is presumed to occur in response to the combined influence of solute-induced variations in fluid density and the upward escape of fluids from the geopressured zone. As a first approximation, fluid density is assumed to be dependent on solute concentration only (Voss, 1984, p. 18)

\[ \rho = \rho_0 + \frac{\partial \rho}{\partial C} \cdot [C - C_0] \]

where

- \( \rho_0 \) = density of fresh water
- \( C \) = solute concentration (kg-NaCl/kg-fluid)
\( C_0 \) = solute concentration of fresh water (=0.)

\( \frac{\partial \rho}{\partial C} = \) rate of change of fluid density with solute concentration

\( = 7.0 \times 10^2 \text{ kg-fluid}^2/\text{m}^3/\text{kg-solute} \)

(Voss, 1984, p. 18)

The viscosity of pure water at 25°C is \( 0.89 \times 10^{-3} \text{ kg.s.m}^{-1} \) whereas the viscosity of a 4 molar NaCl solution at 25°C is \( 1.38 \times 10^{-3} \text{ kg.s.m}^{-1} \) (Phillips et al., 1981). Viscosity changes due to solute concentration are small in comparison to changes that result from temperature alone. A two- to four-fold decrease in viscosity might occur due to temperature changes from the surface to a depth of 3 km (Hanor, 1979, p. 146).

However, even this will not affect fluid flow as significantly as the permeability distribution because permeabilities of sediments vary over six orders of magnitude. Therefore we consider fluid viscosity to be a constant for all simulations following Voss (1981, p. 18)

\[ \mu_0 = 1.0 \times 10^{-3} \quad (\text{kg/m/s}) \]

6. Flow Units

Seven flow units with distinct porosity and permeability values were defined (Fig. 4.6; Table 4.1). Of these, five were defined on the basis of sand percentage for sediments surrounding the dome: more than 50% sand, 25 - 49% sand, 10 - 24% sand (two units), and < 10% sand. The fault was assigned to be part of a sixth flow unit and extended deeper than indicated by the structural cross-section shown by Bennett.
and Hanor (1987, p. 659). In the Welsh study area, it is not possible to
determine the maximum depth of the fault because wells on the north
flank of Welsh Dome do not penetrate deep enough. However, the
existence of faults which extend into the geopressed section has been
documented at the flanks of other salt domes in South Louisiana (Stuart,
1970). Because of our assumption of radial symmetry, Unit 6, which
contains the fault would have a circular fault trace in map plan. Finally,
the salt dome itself was assigned to be a seventh flow unit.

Within flow Units 1 to 4 (sand and shale), porosities range from 0.2
to 0.4, in accord with measurements in whole cores from Calcasieu
Parish, Louisiana, 20 to 40 km from Welsh Dome (Bebout and Gutierrez,
1981). Maximum permeabilities, assumed to be in the horizontal
direction, ranged from $10^{-5}$ darcy for the lowermost geopressed zone
with less than 10% sand to 1 darcy for sediments at the surface (Table
4.1). Our choice of permeabilities for the hydrostatically pressured
interval is consistent with measurements for sediments at a depth of 2 km
surrounding West Hackberry Dome, about 50 km southwest of Welsh
Dome (Fig. 4.8). Our choice of permeabilities for geopressed
sediments is consistent with the range of measured permeabilities for
gelopeared sediments in the Mississippi Fan, $10^{-3}$ darcy to $10^{-8}$ darcy
(Bryant et al., 1985) and the calculations of Bredehoeft and Hanshaw
(1968), Magara (1971) and Bethke (1986).

In our simulations, we assumed that the minimum permeability
was in the vertical direction and that sediments with a lower sand
PERMEABILITY OF SEDIMENTS NEAR WEST HACKBERRY SALT DOME
DARCY
DEPTH RANGE 1670 to 2060 m

Fig. 4.8 Frequency distribution of permeabilities measured from sediment cores at depths of 1670 to 2060 m around West Hackberry salt dome 50 km from Welsh. Most samples lie in the range from $10^{-3}$ to $10^{-2}$ darcy or from $10^{-1}$ to 1 darcy.
percentage were more anisotropic. The ratio of vertical permeability to horizontal permeability ranged from $10^{-2}$ for units with less than 10% sand to 1 for sediments at the surface with more than 75% sand. Unit 6, which contains the fault, was assumed to be isotropic with a permeability of $10^{-2}$ darcy. Porosity and permeability are constant within any one element, and because the number of elements is restricted by constraints on computation time, the smallest element in our simulations was $10^2$ m wide. Our assumption for modeling purposes was that the 'representative permeability' of the fault zone, Unit 6 (Fig. 4.6) is strongly controlled by the fault, even if the fault width is only $10^{-1}$ to 1 m. The salt dome was assigned a permeability of $10^{-12}$ darcy to minimize flow within it and normal to its surface. Measured permeabilities of rock salt range from essentially 0 to $10^{-3}$ darcy (Gloyna and Reynolds, 1961; Bredehoeft, 1988).

A constant value of $1.5 \times 10^{-9}$ m$^2$/sec was used for the mean molecular diffusion coefficient of NaCl in free solution, $D_0$. According to Lerman (1979, p.89), this is the value for 20°C. A constant longitudinal dispersivity of 100 m and lateral dispersivity of 10 m were used for all simulations, values which lie in the 0 to 300 m range for dispersivities calibrated from field tests (Anderson, 1979, p.127).
7. Types of Simulations

We conducted two types of simulations. In the first, we were interested in simulating brine formation by diffusion of dissolved NaCl alone in the absence of significant flow. The rationale here was that if diffusion is not important, then that in itself is evidence of a dynamic subsurface flow regime around Welsh dome. In the second type of simulation, we were interested in the combined effects of solute-induced fluid density gradients and geopressure, and wanted to evaluate the effects of short pulses of upward flow from the geopressed zone. The following subtypes of simulations were conducted with geopressure:

1) Low-permeability geopressed zone (Unit 6). Horizontal permeability was $10^{-5}$ darcy and vertical permeability was $10^{-7}$ darcy.

   a) Fault zone permeability of $10^{-2}$ darcy, isotropic.
   b) Fault zone permeability of $10^{-1}$ darcy, isotropic.
   c) Fault zone permeability of $10^{-2}$ darcy and fault zone extending 300 m below top of geopressed zone (i.e. a little deeper than in Fig. 4.6).
2.) High-permeability geopressed zone. Horizontal permeability was $10^{-3}$ darcy and vertical permeability was $10^{-5}$ darcy.

   a) Fault zone permeability of $10^{-2}$ darcy, isotropic
   b) Fault zone permeability of $10^{-1}$ darcy, isotropic
   c) Fault zone in direct contact with salt dome (with no intervening sediments as in Fig. 4.6). Fault zone permeability $10^{-2}$ darcy, isotropic.

7.1 Boundary and Initial Conditions

For all simulations, nodes within the salt dome were assigned a constant NaCl concentration of 0.3 (30 wt% NaCl), an approximate value for halite saturation at the depth of interest. We assumed that even though salt dissolves with time, the position of the salt-sediment interface does not change with time because new salt is continuously being fed from beneath. At the surface (Fig. 4.6), solute concentration and fluid pressure were assumed to be 0, based on the assumption that the influx of meteoric water or advective-diffusive loss of NaCl would keep surface waters fresh. In fact, a surficial zone of fresh water is widespread in South Louisiana, even where salt domes are common in the subsurface (Rollo, 1960; Winslow et al., 1968). No lateral fluid or solute transport was permitted at the sides. For initial conditions, fresh, hydrostatically pressured pore waters were assumed to exist everywhere at initial time. For the "diffusion-only" simulations, a no-flux condition was assumed for
the base. For simulations involving geopressure, a fluid pressure of 75 MPa was specified at the base (at nodes indicated in Fig. 4.6).

Our assumption of fixed pressures of 75 MPa at the base of the mesh for simulations with geopressure flow are based on fluid pressures observed at a depth of 4 km by Bennett (1985). In a more realistic simulation, fluid pressures at the base of the lower boundary would decrease with time as dewatering of the geopressed zone occurred. Although SUTRA permits the user to specify time-dependent boundary pressures, the depressurizing of the geopressed zone is a complicated problem beyond the scope of this study. The assumption of no lateral fluid or solute transport at the sides of the mesh (Fig. 4.6) is based on the fact that the 'Welsh-type dome' in Fig. 4.6 is actually part of a hypothetical hexagonal array of similar domes such that surrounding domes are spaced 20 km apart from axis to axis.

7.3 "Diffusion only" Simulations

Manheim and Bischoff (1969) observed saline waters above a Gulf Coast salt dome and attributed the formation of these saline waters to diffusion of dissolved NaCl away from the salt domes. Diffusion of dissolved NaCl is an important brine-forming process in certain geologic settings where evaporites occur in the basal part of the section and where solute transport is dominantly vertical (McDuff et al., 1978; Ranganathan and Hanor, 1987). In the first simulation, we wanted to isolate the effects of diffusion alone and neglect the effect of fluid flow
flow, so we set $\partial p/\partial C$ equal to an unrealistically low value, $10^{-3}$ (kg-fluid$^2$/m$^3$/kg-NaCl), in order to minimize gradients in fluid density.

The mean molecular diffusion coefficient of NaCl in sediment was set using a relation which represents a best fit for a variety of deep sea sediments (Berner, 1980, p. 38)

$$D = D^* \sigma^2$$

where $D^*$ = diffusion coefficient of NaCl in free solution at $20^\circ$C

$$(1.5 \times 10^5$ cm$^2$/s, according to Lerman, 1979, p. 89)$$

$\sigma$ = porosity

Since $D^*$ is proportional to absolute temperature in K, $D^*$ at $20^\circ$C is about 70% of $D^*$ at $120^\circ$C, the approximate temperature at a depth of 4 km. Our results should not however be significantly changed, because the functional relation between $D^*$ and sediment diffusion coefficient $D$ used above probably yields a value of $D$ on the high side of what might be observed in partially lithified or clay-rich sediments with high tortuosities.

7.4 Geopressure Flow with Solute-induced Fluid Density Gradients

In these simulations, we set $\partial p/\partial C$ equal to a realistic value of $7.0 \times 10^2$ kg-fluid$^2$/m$^3$/kg-solute (Voss, 1984, p. 18). We induced vertical efflux from the geopressed zone by setting fluid pressures equal to 75
MPa for nodes along the lower boundary (Fig. 4.6). Chemical analyses show that pore waters of geopressed shales from wells only 40 km west of Welsh Dome cluster around 20 g/L TDS (Schmidt, 1973, p. 327). Therefore, in our simulations, we assumed that waters influent into the geopressed zone at nodes indicated in Fig. 4.5 had solute concentration equal to 0.02 or 2 wt% NaCl. For initial conditions, fresh, hydrostatically pressured pore waters were assumed to exist everywhere at initial time. Our results would not be significantly altered by using initial fluid pressures in excess of hydrostatic for the geopressed zone because the geopressure builds up very quickly in the lowermost unit and the pressure field reaches a near-steady state situation in a few thousand years, although the solute concentration field continues to evolve with time.

8. Results

8.1 Role of Diffusion

In the absence of groundwater flow, solute transport occurs by diffusion only. The solute concentration field that results for diffusion after 1 Ma (Fig. 4.9) shows that no brines with more than 5 wt% NaCl are found even a short distance, 0.5 km, away from the dome. The results are not significantly changed even after $10^2$ Ma and indicate that brine formation around salt domes must occur mainly by advection and mechanical dispersion. The observation here that diffusion alone is not an important brine-forming process is largely a consequence of radial
Fig. 4.9 Solute concentration field after 1 Ma for diffusion only, around Welsh Dome. Contours are from $C=0.05$ (5 wt% NaCl) to $C=0.25$ (C=25 wt% NaCl), with a contour interval of 0.05. The solute concentration field is not significantly changed even after $10^2$ Ma, suggesting that diffusion plays a minor role in brine formation around salt domes and that a dynamic subsurface flow regime exists in the Welsh study area.
symmetry and, in itself, is evidence that a dynamic subsurface flow regime must be responsible for brine formation of the sort seen in the Welsh study area. This contrasts with early stages of passive margin evaporite basins where vertical diffusion of dissolved NaCl is an important brine-forming process (Ranganathan and Hanor, 1987).

8.2 Combined effect of geopressure and solute-induced variations in fluid density

**Fluid Pressures**

Vertical cross-sections of the pressure field after $10^{-1}$ Ma show the existence of the geopressed zone below 3 km (Fig. 4.10). The high fluid pressure gradient indicated by close spacing of pressure contours within the geopressed zone is a consequence of the relatively low permeability of the geopressed zone and high fluid pressures, 75 MPa, specified at nodes along the bottom (Fig. 4.6). A pressure drawdown is observed at the top of the geopressed zone near the edge of the dome. We presume this is due to some lateral flow within the geopressed zone near the salt dome and slightly greater fluid effluxes into the hydropressed zone. Away from this pressure drawdown, pressure contours are essentially horizontal. A vertical profile of fluid pressure calculated using SUTRA, at a radius of 5 km (Fig. 4.11) is similar to the profile obtained from shale resistivities for a well 4.5 km north of Welsh Dome (Fig. 4.5). The effect of increasing the horizontal permeability of the geopressed zone to $10^{-3}$ darcy and the vertical permeability to $10^{-5}$
Fig. 4.10 Pressure field after $10^{-1}$ Ma for simulations with low-permeability geopressed zone. Fault zone permeability is $10^{-2}$ darcy, isotropic. Fluid pressure is constant, 75 MPa, at nodes along the bottom in the geopressed sediment. Note the close spacing of pressure contours below 3 km depth corresponding to the geopressed zone. The pressure drawdown in the left corresponds to the edge of the salt dome, where there is some lateral flow within the geopressed zone and where fluid leakage into the overlying hydropressed zone is presumably slightly higher.
Fig. 4.11  Vertical profile of pressures calculated by SUTRA, after $10^{-4}$ Ma, for a radius of 5 km. This profile is similar to the profile of pressures calculated from shale resistivities in Fig. 4.5. Dots correspond to fluid pressures calculated for the case with a low permeability geopressured zone. Dashed line corresponds to fluid pressures calculated for the case with a high permeability geopressured zone. There is very little difference between the two simulations in the hydropressed zone. For the case with a high permeability geopressed zone, the fluid pressure gradient is slightly lower in the geopressed zone.
darcy is to slightly lower the vertical pressure gradient in the geopressed zone, although the general features of the pressure field are similar to the low-permeability case (Fig. 4.11).

Flow Field - Low-permeability Geopressed Zone

When the vertical permeability of the geopressed zone is relatively low, $10^{-7}$ darcy, the flux of fluids vented from the geopressed zone is relatively low and insufficient to force fluid flow up the fault. Instead, the dense brines formed by halite dissolution sink along the fault with a maximum velocity of 0.5 m/yr in the fault zone and spread laterally, overriding the escaping geopressed waters (Fig. 4.12A). As the dense brines sink along the fault, they induce recharge of waters from the shallow subsurface above the dome to the deeper subsurface and induce discharge of waters to the surface away from the dome (Fig. 4.12A). The fault zone is assumed to be isotropic with a permeability of $10^{-2}$ darcy.

Flow Field - High Permeability Geopressed Zone

A very different flow field results when the permeability of the geopressed zone is relatively high ($10^{-3}$ darcy horizontal, $10^{-5}$ darcy vertical). Fluid fluxes out of the geopressed zone are much greater and fluid is focused up the fault with a maximum velocity of 1.1 m/yr (Fig. 4.12B). Inspite of the fact that horizontal permeability is $10^2$ times as great as vertical permeability, much of the water escaping the
Fig. 4.12 Average linear velocities after $10^{-1}$ Ma for two cases. (A) low permeability geopressed zone (horizontal permeability $10^{-5}$ darcy and vertical permeability $10^{-7}$ darcy). (B) high permeability geopressed zone (horizontal permeability $10^{-3}$ darcy and vertical permeability $10^{-5}$ darcy). When the permeability of the geopressed zone is low, dense brines form along the edge of the dome and sink along the fault, shown encircled. When the permeability of the geopressed zone is high, fluids escaping from the geopressed zone are forced up the fault. Lengths of velocity vectors are proportional to $\log_{10} (v_{max})$ where $v_{max}$ is the magnitude of the largest vector. For a scale, the largest vector is plotted in the lower right.
geopressed zone does not actually flow up the fault, but rather flows vertically upwards and around the lower edge of the low-permeability zone represented by Unit 3 at a depth of around 2.5 km (Fig. 4.12B). In this case, the fluid flux from the geopressed zone is sufficiently large that there is net discharge to the surface everywhere (Fig. 4.12B). The fault zone is isotropic and has a permeability of $10^{-2}$ darcy.

**Brine Plume - Low-permeability Geopressed Zone**

Figure 4.13A shows the brine plume that results after $10^{-1}$ Ma assuming the geopressed zone has relatively low permeability ($10^{-5}$ darcy horizontal, $10^{-7}$ darcy vertical) and a fault zone permeability of $10^{-2}$ darcy. Increasing the fault zone permeability to $10^{-1}$ darcy makes the brine plume larger at the same instant in time (Fig. 4.13B). Keeping fault zone permeability at $10^{-2}$ darcy but increasing its depth to within the upper 300 m of the geopressed zone actually permits some brine formation within part of the geopressed zone (Fig. 4.13C).

**Brine Plume - High-permeability Geopressed Zone**

When flow is upward along the fault zone, as in the case of a geopressed zone with relatively high vertical permeability, $10^{-5}$ darcy, the brine plume is formed above the salt dome (Fig. 4.14A). It appears that much of the dissolved salt in the perched brine plume is actually sourced near the edge of the dome and the base of the fault zone. Surprisingly, increasing the fault permeability to $10^{-1}$ darcy does not
Fig. 4.13 Solute concentration fields after $10^{-1}$ Ma for low-permeability geopressed zone. (A) permeability of fault $10^{-2}$ darcy, other permeabilities as in Table 1, distribution of lithologies as in Fig. 4.4. (B) permeability of fault is $10^{-1}$ darcy. Increasing fault permeability allows brines to sink and spread more easily. (C) permeability of fault is $10^{-2}$ darcy, but fault extends into upper 300 m of geopressed zone. The effect of deepening the fault is to actually allow some brine formation in the geopressed zone. The stippled portion corresponds to the observed brine plume bounded by the 160 g/l TDS contour north of Welsh dome in Fig. 4.2, which corresponds to approximately 14 wt % NaCl assuming the TDS content is mainly NaCl.
Fig. 4.14 Solute concentration fields after $10^{-1}$ Ma for high-permeability geopressured zone. The stippled portion corresponds to the observed brine plume bounded by the 160 g/l TDS contour (approximately 14 wt % NaCl) north of Welsh dome in Fig. 4.2. (A) fault zone permeability $10^{-2}$ darcy, isotropic. (B) fault zone permeability is $10^{-1}$ darcy, isotropic. (C) fault zone lies in contact with dome, with no intervening sediments, and fault zone permeability is $10^{-2}$ darcy, isotropic. As a consequence, more salt is dissolved. Note the finger of brine with salt concentrations between 0.05 and 0.10 extending laterally from the top of the dome.
significantly change the rate and geometry of brine plume formation (Fig. 4.14B). The large differences between Figures 4.13A to 4.13C on one hand and Figure 4.14A to 4.14C on the other, suggest that permeability of the geopressed zone is far more important than fault permeability. When the fault zone lies in direct contact with the salt dome, with no intervening sediments as in Figure 4.6, more salt is dissolved and the brine plume is slightly larger (Fig. 4.14C).

8.3 Fluxes From the Geopressed Zone

It is reasonable to ask whether the fluxes of fluid required for formation of perched brine plumes can be sustained for long periods of time, and where the fluids would be sourced from.

Maintaining a constant fluid pressure well in excess of hydrostatic at nodes along a boundary requires that fluids enter the mesh at these nodes. The total fluid flux entering geopressed nodes is balanced by the net discharge at the surface. Using a fluid budget option in SUTRA, discussed by Voss (1984, p. 90 - 92), we calculated to the total fluid influx into geopressed nodes in kg/s, after $10^5$ Ma, and arrived at some surprising results.

When the maximum permeability of the geopressed zone is relatively low ($10^{-5}$ darcy horizontal, $10^{-7}$ darcy vertical), the rate of fluid influx into the geopressed zone is $5$ km$^3/10^5$ yrs, provided that fluid pressure is 75 MPa at the base. Because a near-steady-state pressure
field is achieved at very short times of a few thousand years, the rate of fluid efflux out of the geopressed zone must also be 5 km\(^3/10^5\) yrs. When the permeability of the geopressed zone is relatively high (10\(^{-3}\) darcy horizontal, 10\(^{-5}\) darcy vertical) the fluid efflux from the geopressed zone is 325 km\(^3/10^5\) yrs. Of the 325 km\(^3\) leaving the geopressed zone in 10\(^5\) years, approximately 20% or 50 km\(^3\) escapes up the fault zone.

The large fluxes from the geopressed zone are in large part due to maintaining constant fluid pressures of 75 MPa at 4 km depth. If such large volumes of fluid actually escaped the geopressed zone, the fluid withdrawals would depressurize the geopressed zone. A more realistic model would take into account this depressurizing of the geopressed zone as dewatering proceeds.

A rough estimate of available compaction water can be made from porosity-depth curves for compacted and undercompacted sands and shales in South Louisiana which suggest that a maximum porosity change of 10% results upon compaction (Fig. 4.15). Consider an initially overpressured, undercompacted shale disk 1 km thick and 10 km in radius. The total volume of water available upon compaction is

\[ \Delta \varphi \pi (r_2^2 - r_1^2) \Delta h \]
Fig. 4.15 Porosity of compacted and undercompacted sands and shales in South Louisiana (after Jones, 1975). The maximum porosity difference between compacted and undercompacted sediments is approximately 10%.
where \( \Delta \phi \) = porosity difference between undercompacted and compacted shale = 0.10
\( r_2 \) = radius of shale disk (10 km)
\( r_1 \) = radius of salt dome (1.5 km)
\( \Delta h \) = thickness of geopressured section (1 km)

With these assumptions, a simple volume balance calculation shows that 31 km\(^3\) of pore waters can be expelled by dewatering an undercompacted section that is initially 1 km thick. Thus a 10 km-thick section would have to be dewatered to provide 325 km\(^3\) of water. Many sedimentary basins become geopressed with increasing depth. Although geopressed intervals may extend from a depth of 3 km to the basement in the U.S. Gulf Coast, it is unlikely that the average porosity loss of undercompacted sediments for the entire geopressed section is 10%.

9. Discussion

At this point, it is worthwhile to consider that, the shape of brine plumes formed in our simulations resemble bagels or doughnuts centered about the axis of the salt dome, because of our assumption of radial symmetry. The volume of the simulated perched brine plumes may be larger than that of the observed plume enveloped by the 160 g/l TDS salinity contour in the Welsh study area because the observed plume in the Welsh study area occurs only northwest of the dome and in map plan is elongated rather than being circular (Bennett and Hanor, 1987).
Whereas the perched brine plumes in our simulations occur immediately above the salt dome, the perched brine plume in the Welsh study area occurs about five hundred meters or so above the dome. It is conceivable that our simulations might have matched the observed plume above Welsh dome more closely if we had raised the top of the fault zone or increased permeabilities in sediments overlying the dome. The perched brine plumes of Fig. 4.14C show some semblance of two fingers, like the observed plume in the Welsh study area.

It appears that a substantial portion of the dissolved salt in the perched brine plume is derived from the salty water entering the bottom of the fault zone, but we are unable to determine what fraction of the dissolved salt in the perched brine plume is actually due to this.

There are some other dissimilarities between the observed salinity field in the Welsh study area and the calculated salinity field for the case with a high-permeability geopressure zone. Whereas waters away from the dome are relatively fresh (0 to 5 wt% NaCl) in the simulations of Fig. 4.14, pore waters of sediments below 0.5 km in the Welsh study area have salinities ranging from 10 g/L TDS to 130 g/L TDS (approximately 1 to 12 wt % NaCl). There is actually a decrease in salinity with depth in the geopressured zone. We are unable to account for these features in our simulations and the differences between simulated and observed salinity fields show that the history of subsurface brine formation at Welsh dome has been affected by some very complex factors not taken into account and unknown to us.
A general conclusion we can draw from the simulations presented here is that in the absence of sufficiently large permeabilities for the geopressured zone, any dense brines formed at the top of the dome would sink. In order for a perched brine plume to form, some upward flow is required along the fault zone. If the magnitude of fluxes up the fault zone is sufficiently large, the perched brine plume can form very quickly, in $10^{-1}$ Ma or less.

Two competing effects determine whether perched brine plumes will form: 1) the density-driven sinking of brines off the flank of a dome, the rate of which will be influenced by the permeability structure of flow units surrounding the dome, and 2) possible fluid flow up faults which will be driven mainly by gradients in excess fluid pressure induced by the geopressured zone. Some unanswered questions remain to be addressed in future work. First, is it possible to impede the sinking of dense brines formed at the top of the salt dome by the permeability structure, and still form perched brine plumes by smaller upward fluid fluxes through the fault zone? And second, in the absence of geopressure, how long will it take for perched brine plumes to be dissipated by density-driven flow in the absence of geopressure? The answers to these questions may tell us whether upward flow along faults induced by geopressure is likely to be a currently active process where perched brine plumes are observed over salt domes in South Louisiana. In addition, they may help us constrain the age of formation of perched brine plumes.
10. Conclusions

An anomalous brine plume with more than 160 g/L TDS occurs above and slightly to the north of Welsh Dome in South Louisiana. The formation of this plume was postulated to be due to expulsion of fluids up a fault flanking Welsh Dome, from geopressured sediments below (Bennett and Hanor, 1987). We tested the hypothesis that an pulse of upward flow $10^{-1}$ Ma in duration can create a perched brine plume around a 'Welsh-type salt dome', assuming radial symmetry using a variable density groundwater flow model SUTRA. Although our simulations do not take into account the three dimensional complexity of the Welsh study area and the unknown previous history of brine formation, some general conclusions can be made concerning groundwater flow which results from solute-induced variations in fluid density and from geopressure. We found that anomalous brine plumes can theoretically form in very short periods of time ($10^{-1}$ Ma), but require large effuxes of water from the geopressed zone. Two competing effects determine whether perched brine plumes will form: 1) the density-driven sinking of brines off the flank of a dome, the rate of which will be influenced by the permeability structure of flow units surrounding the dome, and 2) possible fluid flow up fault zones which will be driven mainly by gradients in excess fluid pressure induced by the geopressed zone. The direction of fluid flow in the fault zone is strongly controlled by the permeability of the geopressed zone.
ACKNOWLEDGEMENTS

We express our appreciation to Dave Evans who shared his insights concerning thermohaline convection in salt basins with us and generously allowed us to SUTPLOT, a program used to generate graphic output of SUTRA results. This manuscript has benefited from a review by Jeffrey Nunn. Computer time was provided by the Department of Geology and Geophysics at LSU. We are grateful to Mary Lee Eggart and John Duplechin who prepared some of the illustrations in this paper.

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CONCLUSIONS
Subsurface brine formation is widespread in many evaporite basins. Evidence from Deep Sea Drilling Program site studies suggests that the salinity of pore waters above Miocene evaporites in the Mediterranean and the Red Sea are strongly controlled by diffusion. Calculations done in Chapter 2 show that vertical diffusion of dissolved NaCl can be an important brine forming process in an evaporite basin with a basal evaporite layer. Diffusive loss of NaCl to the surface of the basin may induce dissolution of as much as 250 m of halite provided that the basin-fill is sandy and subsidence rates are relatively slow, 20 to 50 m/Ma. Although young evaporite basins such as the Red Sea and the Mediterranean show classic diffusion-controlled salinity-depth profiles, which are concave upwards, some older evaporite basins such as the Alberta and Michigan basins have salinity profiles which are convex upward. The difference is presumably due to the influence of upward advection and dispersion of dissolved NaCl in slowly flowing groundwater.

In contrast to regions where basal evaporites are horizontal or only gently dipping, advection and dispersion are of much greater importance than diffusion in brine formation near salt domes. Solute-induced fluid density gradients resulting from halite-dissolution are an especially strong driving force for convective overturn of
pore waters around salt domes where active halite
dissolution occurs. The magnitudes of such convective flows
can range from $10^{-2}$ to 1 m/yr even in low-permeability
sediments ($10^{-3}$ darcy). Brine plumes can form quickly by
solute induced convection, and large volumes of salt can be
dissolved in less than 1 Ma. Solute driven convection has the
potential for recharging deep basin sediments with
meteoric waters and inducing sediment diagenesis because
of the significant fluid fluxes that occur over geologic time
intervals especially near the salt dome.

Brine formation around salt domes may also be
influenced by upward flowing waters escaping the
geos pressured zone through fault zones near salt domes. Field
evidence indicates the occurrence of a perched brine plume
above Welsh dome in South Louisiana. Simulations presented
in Chapter 4 suggest that perched brine plumes above salt
domes may form very rapidly, in $10^{-1}$ Ma, provided that
upward flow of geopressured waters occurs near the salt
dome, channeled through a fault near the dome. Two factors
appear to determine whether a brine plume will be perched
above a salt dome or occur around the flanks of the dome: 1)
density-driven sinking of brines off the flank of the salt
dome, the rate of which will be influenced by the
permeability structure of sediments surrounding the dome,
and 2) focused upward flow of geopressured fluids through faults near the dome.

If the permeability of the geopressed zone is low (10^{-5} darcy horizontal, 10^{-7} darcy vertical, in the simulations of Chapter 4, effluxes from the geopressed zone are too small to force fluid flow up the fault and the brine plume sinks. If the permeability of the geopressed zone is larger (10^{-3} darcy horizontal, 10^{-5} darcy vertical) fluids are forced up the fault and perched brine plumes form as ascending geopressed waters flow up the fault. In such a case, effluxes from the geopressed zone must be very large, tens to hundreds of cubic kilometers of fluid.

Two lines of investigation appear to be worthwhile pursuing in future studies. First, detailed mapping of salinity plumes around other South Louisiana salt domes is necessary. The following questions need to be addressed. Are perched brine plumes such as the one above Welsh dome atypical or are they actually common? What is their relation to faults and to the lithologies of sediments surrounding the salt domes? Do they occur in Pleistocene sediments in offshore Louisiana where the Pleistocene section is very thick, or do they occur only in older sediments such as Miocene?
In addition to gathering more field data on subsurface salinity distributions around salt domes, two important questions remain to be answered by numerical modeling. First, if perched brine plumes are a result of episodic dewatering events, how long might they last in the absence of geopressure, before they are dissipated by density-driven flow and dispersion? Also, is it possible to form perched brine plumes with smaller fluxes from the geopressed zone by changing the permeability structure around the salt dome and impeding the sinking of dense brines formed at the top of the dome? The answers to these questions may tell us whether episodic dewatering of geopressed sediments has occurred in the Louisiana Gulf Coast.

Although processes operating over geologic time scales have been of prime interest in this dissertation, a concluding remark is appropriate concerning a hydrogeologic problem that may have a serious impact over man's lifetime - injection of vast amounts of liquid chemical wastes in the subsurface of South Louisiana, fortuitously in sediments around salt domes. In the numerical experiments conducted in this study, permeabilities used were $10^3$ times less than values for shallow sediments and maximum flow velocities ranged from 1 cm/yr to 1 m/yr. It is possible that density-driven groundwater flows may be a thousand times greater in more permeable shallow sediments, particularly if large
lateral salinity gradients exist. It is recommended that the very first step that must be done before considering applications for permits for deep well injection of liquid chemical wastes is to systematically map the lithologic, structural and salinity variations in the area proposed for deep well injection and follow this by a series of controlled numerical model experiments designed to evaluate the impact of various factors on contaminant transport in the subsurface. Of great importance is the hydrodynamics of the unperturbed system as well as that of the system subject to anthropogenic stresses.
Dear Sirs,

I wish to use reprints of the following two papers as part of my dissertation:


2. Ranganathan, V., and Hanor, J.S., (in press), Density-driven groundwater flow near salt domes. Accepted for publication in Chemical Geology (special publication on Fluid-Rock Interactions in the Salt Dome Environment, ed. Posey, H)

Since the bound dissertation will be microfilmed, I request permission to remicrofilm the articles as prepared by Louisiana State University. Thank you.

Sincerely yours,

Vishnu Ranganathan
VISHNU RANGANATHAN was born on September 23, 1955 in Bombay, India. He attended Bombay International High School from which he graduated in 1972. In 1977, he obtained his B.Sc from St. Xavier's college in Bombay. He obtained his M.S. in geology at the University of Cincinnati in 1980 and worked in the petroleum reservoir management division for Cities Service Company in Tulsa, Oklahoma from 1981 to 1983. He started work towards a doctorate in August, 1984, completed his dissertation in September, 1988, and accepted a faculty position in hydrogeology at Indiana University in Bloomington, Indiana.