Salt dissolution and sinkhole formation along the Dead Sea shore

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The formation of sinkholes at the Dead Sea area reflects subsurface cavities formed by salt dissolution. This dissolution is related to the recession of the Dead Sea; the groundwater level and the fresh/saline water interface along the shore decline at a similar rate to the rate of the Dead Sea recession, and brines that used to occupy layers below this interface are flushed out by freshwater. Our finite element modeling shows that dissolution of this salt layer is a plausible mechanism to explain the rapid creation of subsurface holes that collapse and form sinkholes. The positive feedback between the rate of flow, the rate of chemical reaction, and the change in permeability accelerates the dissolution processes and might result in “reactive infiltration instability” which is manifested in “fingers” of cavities, into which fluid is channeled, and salt is dissolved. The spacing between the sinkholes and the rate of their creation is controlled by several factors including properties of lineaments/faults, incoming groundwater flux, the salinity of the incoming groundwater, the rate of dissolution, the effective specific surface area, the permeability of the salt and clay layers, the permeability-porosity relation, the dispersive, and the thickness of the layers. We show that the creation of sinkholes occurs only under specific conditions. These conditions must cause an unstable dissolution front which then causes formation of cavities and eventually sinkholes. The simulations, which utilized the best estimated parameters of the studied area, yield results that are similar to those exhibited in the field.


1. Introduction

Formation of sinkholes is a well known phenomenon, occurring in many parts of the world (Spain, United States, Italy, Thailand, and more). The mechanism that is responsible for the appearance of sinkholes is dissolution of soluble rocks and creation of subsurface cavities [Martinez et al., 1998; Galloway et al., 1999; Gutierrez and Cooper, 2002]. The dissolved rock is usually carbonate [Tihansky, 1999], whose rate of dissolution is rather slow, or salt whose rate is much higher. Salt dissolution is a very fast process that alters the hydraulic properties of both the porous media and the pore fluid and can lead to a reaction-infiltration feedback loop [Ortoleva et al., 1987].

Sinkholes along the western shore of the Dead Sea (Figure 1) have become a major concern in the last decade with the appearance of hundreds of sinkholes [Arkin and Gilat, 2000; Abelson et al., 2003]. They indicate the creation of subsurface cavities due to dissolution of layers of salt by fresh groundwater [Abelson et al., 2006].

1.1. Reactive Infiltration Instability

The kinetics of salt dissolution depends on the rate of solute transport and on the rate of the chemical reaction between the solution and the solid phase: high rates of solute transport and chemical reaction cause faster dissolution [Steefel and Lasaga, 1994]. In addition, the dissolution of salt increases the porosity and the permeability of the rock, thereby increasing both the rate of solute transport and the rate of the chemical reaction. This positive feedback between the rate of flow, the rate of chemical reaction, and the change in permeability accelerates the dissolution processes and results in “reactive infiltration instability” [Ortoleva et al., 1987; Aharonov et al., 1997]. This instability is manifested in “fingers” of cavities, into which fluid is channeled and salt is dissolved. The reactive infiltration instability has been studied extensively in relation to acidization that causes dissolution in carbonate rocks [Daccord and Lenormand, 1987; Fredd and Fogler, 1998], salt deposits [Béki et al., 1995] geochemical reaction fronts [Ortoleva et al., 1987; Steefel and Lasaga, 1994; Raffensperger and Garven, 1995; Bolton et al., 1997] and melt extraction from the mantle [Aharonov et al., 1997]. In this paper we study the processes and parameters that control salt dissolution that lead to the formation of sinkholes along the Dead Sea shore.

1.2. Hydrogeology of the Dead Sea Area

The Dead Sea basin is one of the largest pull-apart basins on Earth and is filled with Quaternary sediments. The shallow section of the basin fill is composed of gravel with an intercalation of salt and clay sediments that divide the gravel into few subaquifers [Yechieli, 2000]. The salt layer...
is usually separated from the gravel by clay layers. The hydraulic head in the deeper confined aquifer is higher than the hydraulic head in the shallow phreatic aquifer. Freshwater enters the aquifers from the Judean aquifer at the west, and flows eastward to the Dead Sea while mixing with brines. The diversion of freshwater from the Jordan River since the 1960s, by both Jordan and Israel, has reduced the flow of water into the Dead Sea so that the volume of water evaporated is now greater than the volume of water input. As a result, the sea level is rapidly dropping at a rate of almost 1 m yr\(^{-1}\) (Figure 2). The groundwater level and the fresh-saline interface response to the Dead Sea recession is a

**Figure 1.** Location map showing the sinkhole sites along the Dead Sea.

**Figure 2.** Dead Sea level (solid line) and number of sinkholes at the western shore of the Dead Sea (dashed line).
continuous decrease whose magnitude is a function of the distance from the shoreline [Yechiel, 2000].

1.3. Sinkholes Along the Dead Sea Shore

The formation of sinkholes at the Dead Sea area has a strong connection to the recession of the Dead Sea level: The sinkholes are mainly located in areas that were submerged before the water recessed and the rate of their formation accelerates while the Dead Sea level continues to recede (Figure 2). The sinkholes began to appear about 20 years after the Dead Sea level started to decline. Because of the recession of the Dead Sea level, and the decline in the fresh/saline water interface along the shore, brines that used to occupy layers below this interface are now being flushed out by freshwater (Figure 3). If these layers contain salt, the freshwater dissolves the salt and creates voids. The collapse of these voids makes the sinkholes. Abelson et al. [2003] shows that the sinkholes appear along lineaments and, based on seismic reflection profiles and interferometric synthetic aperture radar measurements, they concluded that the sinkholes track young fault systems. The faults serve as conduits, channeling freshwater from the deeper aquifer to the shallower one, promoting the development of sinkholes. There are several other options for fresh groundwater to come into contact with the salt layer, such as the case where the clay layer below the salt is missing, or at the edge of the salt layer where there may be a direct contact between gravel and salt. In the present paper, one of the main possibilities, the existence of fault affecting the groundwater flow, is examined.

The goal of this study is to provide a quantitative description of this conceptual model and to constrain the range of physical parameters that control the temporal and spatial characteristics of subsurface cavity development.

2. Formulation of the Problem

The governing equations for reactive flow are conservation of fluid mass and conservation of solute mass. The flow equation derived from Darcy’s law and the conservation of fluid mass for isothermal, variable density conditions is

\[
\nabla \cdot \left[ K \rho \frac{\partial h}{\partial t} \right] = \rho S_s \frac{\partial h}{\partial t} + RA (C - C_{sat})
\]

where \( K \) is the hydraulic conductivity tensor (L T\(^{-1}\)), \( h \) is the freshwater hydraulic head (L), \( \mu_r = \mu_0/\mu \) is the relative viscosity (dimensionless), \( \mu \) is the viscosity of water (M L\(^{-1}\) T\(^{-1}\)), \( \mu_0 \) is the water viscosity at standard state (M L\(^{-1}\) T\(^{-1}\)), \( \rho \) is the fluid density (M L\(^{-3}\)), \( \rho_s = (\rho - \rho_0)/\rho_0 \) is the relative density (dimensionless), \( \rho_0 \) is the density at standard state (M L\(^{-3}\)), \( S_s \) is the specific storage coefficient (L\(^{-1}\)), \( R \) is the reaction rate constant of the soluble component (M L\(^{-2}\) T\(^{-1}\)), \( A \) is the effective specific surface area (L\(^{-1}\)), \( C \) and \( C_{sat} \) are the actual concentration and the saturation concentration of the soluble component in the fluid, respectively (dimensionless).

Solute transport through porous media is controlled by advection, dispersion, and dissolution/precipitation. The conservation of solute mass may be written as

\[
\frac{\partial (\phi C)}{\partial t} = \nabla \cdot \left[ D \nabla (\phi C) \right] - q \cdot \nabla (\phi C) - \frac{RA\phi}{\rho} (C - C_{sat})
\]
where $\phi$ is the porosity (dimensionless), $D$ is the dispersion-diffusion tensor ($L^2 T^{-1}$), and $q$ is the average linear velocity of the fluid ($L T^{-1}$). The solute mass exchange between the solid and fluid in a porous media modifies the properties of both the solid and the fluid. The fluid’s density and viscosity are related to the solute concentration through equations of state [Watson et al., 1980; Phillips et al., 1981]. The Dead Sea brine has a unique composition and does not follow exactly these general equations of state which are derived for NaCl solutions. However, we use the equations of Watson et al. [1980] and Phillips et al. [1981] because of the lack of a complete set of equations of state for the Dead Sea brine. At atmospheric pressure and temperature of 25°C, the calculated density and viscosity are different from that of the Dead Sea density and viscosity by 90% and 80%, respectively (N. Weiszbrod, unpublished data, 2005). Compared to the uncertainty in the permeability values, these differences are insignificant. The solute added to the fluid phase is subtracted from the solid phase increasing the porosity of the medium. This change in porosity will increase the permeability as [Bernabé et al., 2003]

$$k = k_0 \left( \frac{\phi}{\phi_0} \right)^n$$

(3)

where $k_0$ and $\phi_0$ are the initial permeability and porosity, respectively, and $n$ an empirical variable. Laboratory experiments show that the exponent, $n$, varies by order of magnitude, in most cases in the range of 2–3, but could be nearly infinite [McCune et al., 1979; Noiriel et al., 2004]. Bernabé et al. [2003] showed that this relationship depends on the process taking place, i.e., elastic or plastic compaction, diagenesis, dilatancy, dissolution, precipitation. This exponential relationship implies that in places where dissolution starts the permeability increases significantly, channeling more fluid into the dissolved areas, thereby accelerating the process. This is an unstable process that is controlled by the porosity-permeability relation and by two dimensionless numbers: Péclet (Pe) and Damköhler (Da):

$$Pe = \frac{ql}{D} = \frac{\text{flow velocity}}{\text{dispersion velocity}}$$

(4)

$$Da = \frac{\text{reaction rate}}{q} = \frac{\text{flow rate}}{\text{reaction rate}}$$

(5)

where $l$ is a characteristic length. For $Da$ larger than 1, the reaction front tends to destabilize, creating channel like formations [Steefel and Lasaga, 1994]. The shape of these features depends on the Pe number. For large Pe numbers, the channels become long and narrow [Chen and Ortoleva, 1990].

[10] Once the porosity and permeability increase to very high values, as a result of the dissolution, Darcy’s law does not hold anymore due to the development of turbulent flow [Bear, 1972]. Reynolds number is used to distinguish between laminar flow occurring at low velocities and turbulent flow:

$$Re = \frac{ql}{\nu} = \frac{\text{inertial force}}{\text{viscous force}}$$

(6)

where $\nu$ is the kinematic viscosity of the fluid. Darcy’s law is valid only for linear-laminar flow indicated by $Re < 10$. For $10 < Re < 100$ the flow is laminar but nonlinear, and for $Re > 100$ the flow is turbulent [Bear, 1972]. The Navier-Stokes equation accounts for all regimes of flow adequately describing the transition from porous medium flow to pure fluid flow. A solution for this equation would require extremely fine discretization to model the details of the turbulence [Zienkiewicz and Taylor, 2000]. Brinkman [1947] suggests that for the nonlinear laminar regime the inertial term can be omitted from the Navier-Stokes equation, and still account for the dynamic viscosity. Ormond and Ortoleva [2000] use the Brinkman’s equation to model dissolution and show different dissolution patterns than a model that uses Darcy’s law. In this work we model fluid flow, solute transport, and dissolution within the porous media and not within the cavities. Therefore, once the porosity of an element is higher than a critical value we assume that its permeability is high enough to cause negligible head gradient within the cavity compared to the head gradient in the porous media. This assumption allows us to only use Darcy’s law.

3. Model Description

[11] Equations (1) and (2) are solved using a Galerkin finite element technique with linear shape functions applied over triangular elements for two-dimensional cross sections. The top boundary is a free surface that represents the water table (pressure of 0) and its position is calculated at every time step as described by Neuman and Witherspoon [1971]. Because the water table position is changing in a stationary Cartesian coordinate plane (Eulerian formulation), it is necessary to describe the position of the water table within the elements. Every element is numerically integrated over seven Gauss points at which the pressure is calculated to find the actual position of the water table.

[12] The formation of cavities at the Dead Sea area is studied using two perpendicular sets of two-dimensional simulations (Figures 4a and 4b). In the first set of simulations, a hydrological system of two subaquifers is modeled (Figure 4a). In this simulation, the east boundary represents the Dead Sea water column which has receded by 22 m since 1970. On the basis of a rough estimate of the groundwater balance along the Dead Sea shore, a flux of 10 m² (m yr)⁻¹ is assigned to the west boundary representing the input of water from the Judean aquifer west of the cross section. Because we did not want the west boundary to be close to the area of interest, we continued the cross section to the west for another 1000 m that are not shown. This allows the salinity on the west side to vary during the simulation. No flow boundary conditions are assigned at the bottom boundary where a clay layer divides these upper layers from deeper ones. Steady state conditions for 1970 were obtained by assigning salinity of the Dead Sea for the entire cross section initially, and then flushing the cross section with fresh water from the west, forming a fresh/saline water interface, until no changes in salinity and hydraulic head were observed. Hydrological properties of the different units are shown in Table 1. The reaction rate constant is $2 \times 10^{-4}$ kg m⁻² s⁻¹ ±15% [Alkattan et al., 1997] and the equilibrium concentration of the soluble
component in the fluid is 0.275 g g$^{-1}$. The effective specific surface area is not known and could vary by order of magnitudes. Renard et al. [1998] reported a value of 350 m$^{-1}$ and Bercovici et al. [2001] argue that it can be as high as 10$^6$ m$^{-1}$. Ni and Beckermann [1991] show that the effective specific surface area is related to the porosity as $A \sim \phi (1 - \phi)$. In our simulation the effective specific surface area is set to 350 m$^{-1}$.

The second cross section represents a section of the fault plane where the fault crosses the salt and clay layers (Figure 4b). The cross section is initially saturated with the Dead Sea brines and is then flushed by freshwater from the bottom boundary. Boundary conditions as well as fluid and solid properties are changed between simulations as will be discussed later.

4. Results

4.1. Cross Section Perpendicular to the Fault Plane

The Dead Sea level has been dropping rapidly since the 1970s after 10 years of an almost steady water level (Figure 2). Therefore we assume that in 1970 the system was at steady state conditions. The recession of the groundwater level and the fresh/saline interface as the Dead Sea level fell is shown in Figure 2.

**Figure 4.** Finite element meshes and boundary conditions for two cross sections: (a) perpendicular to the fault and (b) along the fault.

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**Table 1. Simulation Parameters**

<table>
<thead>
<tr>
<th>Gravel$^a$</th>
<th>Clay</th>
<th>Salt</th>
<th>Fault</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal Permeability, m$^2$</td>
<td>Vertical Permeability, m$^3$</td>
<td>Porosity, $%$</td>
<td>Specific Storage, m$^{-1}$</td>
</tr>
<tr>
<td>$5 \times 10^{-12}$</td>
<td>$5 \times 10^{-14}$</td>
<td>20</td>
<td>$10^{-3}$</td>
</tr>
<tr>
<td>$10^{-17}$</td>
<td>$10^{-18}$</td>
<td>10</td>
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<td>$10^{-18}$</td>
<td>$10^{-13}$</td>
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<tr>
<td>$10^{-12}$</td>
<td>$10^{-17}$</td>
<td>20</td>
<td>$10^{-3}$</td>
</tr>
</tbody>
</table>

$^a$Based on results of pumping tests along the Dead Sea shore [Hirsch, 1975; Wollman et al., 2003].
Sea level declines, along with the groundwater flow velocities, are shown in Figure 5. Freshwater enters the cross section from the left boundary and flows eastward on top of the Dead Sea brines, eventually discharging at the Dead Sea shore. A groundwater flow of 10 m yr$^{-1}$ is assigned to the left boundary (not shown), which is located 1000 m left of the edge shown in Figure 4a. At the left edge of the cross section shown in Figure 5a, fluid velocity is determined to be between 20 and 30 m yr$^{-1}$. This is about twice the flux assigned to the left boundary but taken over an area half the size, thus conserving the total fluid mass.

The Dead Sea brines enter the cross section from the bottom right boundary, migrate upward in the fault and around the salt and clay layers, mix with freshwater, and finally discharge back to the Dead Sea under the Dead Sea shore. Groundwater velocities are negligible within the salt and clay layer because their permeabilities are significantly lower than the permeability of the gravel (Table 1). As the Dead Sea level goes down, the groundwater level drops and the fresh/saline water interface moves eastward and drops (Figure 5b–5e). Brines no longer enter from the right boundary. Instead, the brines are flushed back into the Dead Sea because the hydraulic head at the Dead Sea drops. This is true in both subaquifers, below and above the salt layer. Initially, the water flowing up the fault is brine. With the recession of the water level and the fresh/saline water...

Figure 5. Groundwater level (top solid line), fresh/saline water interface (thick solid line), and groundwater velocity arrows evolution. Velocity arrows magnitude is proportional to their length. The reference velocity vector has magnitude of 100 m yr$^{-1}$. 
interface, the water flowing up the fault become less saline until finally, its salinity is half the salinity of the Dead Sea brine.

16 The evolution of porosity and hydraulic head in the area where the fault crosses the salt layer is shown in Figure 6. When the porosity is above 40%, the element becomes a cavity and we do not calculate the fluid velocity. We ran several simulations and found that results are not sensitive to this value. The simulated cavity grows upward and then expands laterally. The difference in the calculated freshwater hydraulic head between the upper and lower aquifers is about 3 – 4 m, which is consistent with the hydraulic head measured in boreholes. This difference is controlled by the permeability of the clay layers, which do not dissolve, and therefore remains at the same magnitude even when the cavity extends across the entire depth of the salt layer.

17 From this simulation we see that the flow within the fault where it intersects the salt and clay layer is always upward (Figure 5). Initially, Dead Sea brines flow up the fault (Figure 5a), with time the water salinity decreases within the fault until the salinity of the water is half of the salinity of the Dead Sea (Figure 5e). The model for the simulation along the fault is designed to model groundwater flow up the fault where it crosses the salt and clay layers (Figure 4b).

4.2. Cross Section Along the Fault Plane

18 The boundary conditions for the simulations along the fault plane are assigned based on the results from the simulations perpendicular to the fault plane. Fluid flux across the fault where it intersects the salt and clay layers is <1 m² (m yr)⁻¹ and the difference in hydraulic head between the top and bottom of the salt layer is ~4 m. We create a reference simulation which is used to demonstrate the process of cavity creation as well as to use in a sensitivity analysis of the system. The characteristics of this reference simulation are as follows: the flux of fresh water from the lower boundary is set to 0.2 m yr⁻¹, the hydraulic head at the top boundary is constant, the initial conductivity and porosity of both the salt and clay are 0.1 m yr⁻¹ and 0.1, the diffusion coefficient is 0.001 m² yr⁻¹, there is no dispersion, the dissolution reaction rate is 0.0002 kg m⁻³ s⁻¹ [Alkattan et al., 1997], the saturation state of salt is 0.275 g g⁻¹, the effective specific surface area is 350 m², and the permeability porosity relation is \( k = k_0 (\phi/\phi_0)^3 \).

19 From this simulation we see that the process that forms sinkholes begins as a reaction front and becomes unstable soon after the fault begins to be flushed with freshwater (Figure 7a). Thereafter, each cavity becomes unstable and starts to produce subcavities (Figure 7b). With time, these subcavities rejoin to form larger cavities.
(Figures 7c and 7d). Some cavities cease to develop when water is channeled to a nearby cavity (Figures 7d–7f). The cavities’ tips develop in a dome shape until they reach the top clay layer, where they then expand laterally, taking on the shape of an overturned cone. A dome is mechanically much more stable than an overturned cone due to the intrinsic stability of the arch form. Furthermore, salt is much stronger than clay and gravel, and above the top clay layer there is only 20 m thick gravel layer which cannot hold large loads. Therefore we assume that when the cavity has a dome shaped tip and is within the salt layer, it is stable mechanically. Once the cavities reach the top clay layer and change their shape, they become unstable and collapse to form sinkholes.

\[20\] In the following, we perform a sensitivity analysis and show the significance of different parameters in controlling fluid flow and dissolution patterns.

### 4.2.1. Incoming Flux

\[21\] The rate of sinkhole formation and the total number of sinkholes strongly depends on the groundwater flux entering the fault zone from the lower boundary (Figure 8). At low fluxes (0.1 m yr\(^{-1}\)) the sinkholes start to form at later stages and their rate of formation decreases (slopes in Figure 8). At fluxes greater than 0.5 m yr\(^{-1}\), the rate of sinkhole formation does not change much because it is limited by the dissolution chemical reaction rate.

\[22\] At initial stages the reaction front advances in a constant pace until fingers are formed. Thereafter, the pace at which the fingers advance accelerates while the pace at which the main reaction front advances slows down. The rate of sinkhole formation is high at high fluxes (>0.5 m yr\(^{-1}\)) because the fingers are created close to the top clay layer and have a very short distance to advance before they create sinkholes. At low fluxes (0.1 m yr\(^{-1}\)), the fingers are created far below the top clay layer. Small differences in the velocities at which the fingers advance result in large differences in their arrival time to the top clay layer where the collapse of the sinkhole occurs. Therefore the slopes in Figure 8 are steep at high fluxes and moderate at low fluxes. At intermediate fluxes (0.2–0.3 m yr\(^{-1}\)), sinkhole formation is initially moderate and accelerates after a few years. For example, at a flux of 0.2 m yr\(^{-1}\), the first three sinkholes are formed in two years, and the following six sinkholes are formed in 2.5 years. Sinkhole formation accelerates from 1.5 sinkholes per year in the first two years, to 2.5 sinkholes per year in the following two and half years. The reason for this is that fingers are formed during the migration of the main reaction front upward. Fingers that are formed initially far away from the top clay layer get to the clay layer at different times depending in their velocity. Later, when the main reaction front gets closer to the top clay layer, the distance that the fingers cross is smaller and they arrive almost at the same time. The deceleration in sinkhole formation at final stages reflects the arrival of the main reaction front to the top clay layer and the end of dissolution.

\[23\] Fluxes below 0.1 m yr\(^{-1}\) and above 1 m yr\(^{-1}\) do not result in sinkholes but instead in a long horizontal flux migrating upward and eventually forming a tunnel (Figure 9). However, the propagation of the reaction front is different at high and low fluxes; at a high flux, the reaction front is stable and forms a straight line (Figure 9a), whereas at a low flux, the reaction front is unstable and forms fingers that are short and wide and do not form discrete sinkholes.
At intermediate fluxes (0.2–0.5 m yr\(^{-1}\)), sinkhole creation accelerates 1–3 years after the first sinkhole was created.

### 4.2.2. Water Level Recession

[24] The Dead Sea level and groundwater levels have been receding since the 1970s. This is simulated by reducing the hydraulic head at the top boundary by a rate of 1 m yr\(^{-1}\), which is the rate of the Dead Sea recession. Simulated sinkhole evolution over time is shown in Figure 10. When the flux from the lower boundary is high, the sinkhole evolution is similar to the simulation with constant hydraulic head at the top boundary (Figure 8). When the flux from the lower boundary is low, the influence of reducing hydraulic head at the top boundary is significant. The water level recession adds another driving force for fluid flow, in addition to the injection of freshwater at the lower boundary. These two driving forces act in different ways on the fluid flow. The freshwater flux prescribed at the lower boundary, displaces brines, dissolves salt, and create cavities. The recession in water level at the top boundary "pumps out" brines from the cross section and changes the hydraulic head near the top boundary, far away from the less saline water at the bottom part of the cross section. By the time the top boundary starts to influence the bottom part of the cross section, it has already created a large upward hydraulic gradient that sucks the less saline water upward, enhancing fingering. For example, at flow equal to 0.1 m yr\(^{-1}\), sinkholes start to form at 17 years and then there is a fast growth. Without water recession at the top boundary (Figure 8) the onset is 21 years and sinkhole growth is slow. When the incoming flux from the lower boundary is higher, the brines are flushed away by this flux in less than 14 years, too fast for the water level drop at the top boundary to influence the process.

### 4.2.3. \(K-\phi\) Relationship

[25] The permeability-porosity relationship is a crucial factor in determining dissolution and fluid patterns (equation (3)). A simulation where the exponent \(n\) is equal to 8 is shown in Figure 11a. This accelerates dissolution and creates long and narrow fingers.

[26] Laboratory work relates porosity to permeability in the direction of flow. Permeability perpendicular to the flow direction might not evolve in the same matter. A simulation where the exponent \(n\) is equal to 8 for the permeability in the direction of the flow and to 2 for the direction perpendicular to the flow is shown in Figure 11b. The permeability parallel to the flow increases faster than the permeability perpendicular to the flow causing fluid flow and dissolution to be preferably upward creating, longer and narrower fingers.

### 4.2.4. Boundary Conditions at the Lower Boundary

[27] The lower boundary, which is prescribed by a flux boundary condition, can also be prescribed by a hydraulic head. Fluxes at the lower boundary in a simulation where the hydraulic head at the lower boundary is 3m higher than at the top boundary are compared to a simulation with the same conditions but with no dissolution in Figure 12. Initially, while the fresh water flows through the clay, both simulations give the same result. Once the fresh water reaches the salt layer after three years, dissolution begins,

**Figure 9.** Porosity evolution of the fault for (a) low incoming flux and (b) high incoming flux.

**Figure 10.** Sinkhole creation as a function of the incoming flux from the lower boundary with groundwater level recession.

**Figure 11.** Porosity evolution of the fault (a) for \(n = 8\) and (b) for \(n = 8\) parallel to flow direction and \(n = 2\) perpendicular to flow direction.
and the flux in the simulation where dissolution is taking place is higher than the flux where dissolution is not taking place. While dissolution increases salt permeability, the flux through the cross section increases. After 14 years, fingers connect the two clay layers so that the hydraulic gradient is minimal at the salt layer and maximum at the clay layers.

4.2.5. Dispersivity

Dispersivity changes by orders of magnitude depending on the field scale [Gelhar et al., 1992; Neuman and Di Federico, 2003]. When the scale of observation is of meters the longitudinal dispersivity is 0.01–0.1 m and the transverse dispersivity is 0.001–0.01 m. Dispersivities multiplied by the groundwater velocity give the dispersion coefficients. Figure 13 shows the effect of dispersivity. Including the dispersivity decreases the Peclet number, which results in wider fingers. Having different longitudinal and transverse dispersivities also increases the anisotropy, resulting in more fingers. In these simulations the typical fluid velocity is 0.1–1 m yr$^{-1}$, so that the coefficient of dispersion has the same order of magnitude as the molecular coefficient (0.001 m$^2$ yr$^{-1}$).

4.2.6. Initial Permeabilities

The permeability of the lithological units controls the fluid velocity and changes both the Peclet and Damköhler numbers. Therefore the effect of simulations with different permeabilities is similar to that of different incoming fluxes as indeed was expected. However, another complication is introduced when the initial permeabilities of the salt and clay layers are varied independently. The permeability of the clay controls the flux entering the cross section, or the hydraulic head, depending on the boundary condition employed. The initial permeability of the salt has no importance once a finger breaks through the salt layer connecting the two clay layers.

4.2.7. Salt Layer Thickness

As expected, the rate of sinkhole formation depends strongly on the thickness of the salt layer. Figure 14 shows sinkhole evolution as a function of salt layer thickness. The dissolution reaction front is advancing at a typical velocity and the time that is required for the front to get to the top clay layer is a function of the velocity and the thickness of the salt layer. Without these clay layers, once a cavity breaks through the salt layer, it would channel all the water and stop the evolution of the rest of the cavities. The clay layers buffer the fluid flow and do not allow fast fluid flow through the cavity. As a result, other cavities continue to develop.

5. Discussion

The model described in this paper relies on two observations: (1) most sinkholes align in straight lines and (2) in most sinkhole sites, the salt layer is separated from the gravel aquifer by impermeable clay layers. Abelson et al. [2003] suggest that the lines reflect young, permeable faults that serve as fluid conduits. As the Dead Sea level drops, freshwater dilutes and flushes the brines upward through the faults back to the Dead Sea. The fault disrupts the clay layers and allows groundwater flow into and past the salt. For this to occur, the faults must be active, to keep fault permeability high. Alternatively, the clay layers may be noncontinuous, so that in places the salt layer is directly attached to the gravel aquifer. In this case, dissolution occurs in these isolated places and not within a fault. We also assumed that the lower subaquifer is open to the Dead Sea (Figure 4a) resulting in a large volume of groundwater flow to the Dead Sea through this aquifer. In reality, the clay and salt layers may dip down toward the Dead Sea, sealing off groundwater flow into the Dead Sea. As a result, the groundwater flow through the fault would be larger than what we calculated and the rate of sinkholes creation would be significantly higher.

The nonuniqueness of the problem and the great variability between sinkhole sites makes it hard to predict sinkhole evolution. There are other possible scenarios. The recent study shows a realistic possibility in which upward flow occurs mainly through a fault. The hydrologic and kinetic properties of any specific site are poorly constrained and could range over a wide spectrum. Nevertheless, we are able to find reasonable values for fingering and sinkhole evolution. The flux from the lower boundary when the lower boundary condition is constant hydraulic head: with dissolution (curve a) and no dissolution (curve b).
formation. We explain the 20 year delay in the onset of sinkhole formation after the recession of the Dead Sea began: it takes 20 years for the freshwater to flush the brines all the way to the salt layer and then to dissolve the salt to create a cavity through the salt up to the shallower clay layer. The sinkholes are initially widely distributed, and then new sinkholes are created in between them. During the migration of the unstable front new fingers are created and existing fingers join each other. At intermediate fluxes, sinkhole formation accelerates as actually observed at the Dead Sea area.

In general, unstable reaction fronts with finger ing occur at high Damköhler numbers \cite{Daccord and Lenormand, 1987; Steefel and Lasaga, 1994}. At high Peclet numbers the fingers are long and narrow and at low Peclet numbers they are short and wide \cite{Ortoleva et al., 1987; Aharonov et al., 1997}. Fresh groundwater flux entering the cross section controls both the Peclet and Damköhler numbers (equations (4) and (5)). Low incoming flux means high Damköhler number and low Peclet number. Therefore it is expected that well developed fingers that cause sinkholes will occur at intermediate fluxes, where both Peclet and Damköhler numbers are high. Indeed, our results show that sinkholes develop only at fluxes between 0.1 and 1 m yr\(^{-1}\). At higher and lower fluxes the dissolution results in elongated channels (Figures 8 and 9). At high fluxes, where the Damköhler number is small and the Peclet number is high the reaction front is stable and there is no fingering. At these high fluid velocities, dissolution cannot keep up with the fluid flow, so the pattern of flow is not controlled by dissolution. At very low fluid velocities, dissolution is much faster than fluid flow and fingering occurs. However, when fluid flow is so slow that salt diffuses from the finger walls at a similar rate, widening the fingers, it results in essentially a straight line that does not produce sinkholes. The dissolution rate depends on the effective specific surface area \((A)\), the dissolution rate constant \((R)\), the equilibrium concentration of the soluble component in the fluid \((C_{\text{sat}})\), and the concentration in the water entering the system \((C)\) (see equation (2)). High values of \(R\) and \(A\) mean a high Damköhler number and unstable dissolution. Although not included in this dimensionless number, the difference \((C-C_{\text{sat}})\) has a similar effect.

As stated before, there are other possible scenarios that can be simulated. Such scenarios include a case where there is no fault or where the fault’s effect is not that strong. The connection of the lower subaquier to the Dead Sea is not clear, and the hydraulic parameters could change by orders of magnitude between sites. However, according to the field data, the case of a permeable fault given here is the most reasonable one to describe the actual conditions in many sites.

The mechanism that is responsible for the reactive infiltration instability (i.e., fingering) in the porous media is dependent on the fact that permeability is related to porosity. Once porosity is increased by dissolution, fluids are channeled into the dissolved sections and accelerate the process. In other words, the heterogeneity created by dissolution magnifies this instability. If initially the porous media is heterogeneous, this instability increases, the reaction front is not as smooth, and the finger growth rate is faster \cite{Ormond and Ortoleva, 2000}. In fact, in many cases in order to increase fingering, nonuniformity was assigned \cite{Ortoleva et al., 1987; Renard et al., 1998; Ormond and Ortoleva, 2000}.

6. Conclusions

Reasonable combination of various model parameters including flux, hydraulic head change, porosity-permeability relation, and salt layer thickness can reproduce the observed rate and pattern of sinkhole formation. This strongly supports the mechanism of salt dissolution as a major process that controls sinkhole formation. Dissolution of a salt layer as a result of water level recession is shown to be a plausible mechanism to explain the fast creation of sinkholes at the western shore of the Dead Sea over the past 30 years. The recession of the Dead Sea level causes a recession in the groundwater level and in the fresh/saline water interface. The Dead Sea brines are flushed with freshwater that dissolves salt layers and creates sinkholes. The spacing between the sinkholes and the rate of their
creation is controlled by the properties of lineaments/faults, the incoming groundwater flux, the salinity of the incoming groundwater, the rate of dissolution, the effective specific surface area, the permeability of the salt and clay layers, the permeability-porosity relation, the dispersivity, and the thickness of the layers. We show that the creation of sinkholes occurs only under specific conditions that cause an unstable dissolution front.

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