Transient groundwater-lake interactions in a continental rift: Sea of Galilee, Israel

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ABSTRACT

The Sea of Galilee, located in the northern part of the Dead Sea rift, is currently an intermediate fresh-water lake. It is postulated that during a short highstand phase of former Lake Lisan in the late Pleistocene, saline water percolated into the subsurface. Since its recession from the Kinarot basin and the instantaneous formation of the fresh-water lake (the Sea of Galilee), the previously intruded brine has been flushed backward toward the lake. Numerical simulations solving the coupled equations of fluid flow and of solute and heat transport are applied to examine the feasibility of this hypothesis. A sensitivity analysis shows that the major parameters controlling basin hydrodynamics are lake-water salinity, aquifer permeability, and aquifer anisotropy. Results show that a highstand period of 3000 yr in Lake Lisan was sufficient for saline water to percolate deep into the subsurface. Because of different aquifer permeabilities on both sides of the rift, brine percolated into aquifers on the western margin, whereas percolation was negligible on the eastern side. In the simulation, after the occupation of the basin by the Sea of Galilee, the invading saline water was leached backward by a topography-driven flow. It is suggested that the percolating brine on the western side reacted with limestone at depth to form epigenetic dolomite at elevated temperatures. Therefore, groundwater discharging along the western shores of the Sea of Galilee has a higher calcium to magnesium ratio than groundwater on the eastern side.

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INTRODUCTION

Rift lakes are dynamic environments with rapid and large hydrologic fluctuations that are accompanied by drastic variations of water salinity, surrounding groundwater heads, and sedimentation patterns (Crossely, 1984; Cohen, 1989; Scholz et al., 1990). An example is Lake Malawi, where water level was 200–300 m lower than the present level between 40 and 28 ka, during which the lake had a different chemical composition (Finney et al., 1996). These rapid fluctuations reflect the complex interplay between climate, which affects the volume of water flowing into the lake, and the rate of evaporation and tectonics, which affects the positioning of pathways and barriers in the basin-to-water input and output (Street-Perrott and Harrison, 1985; Benson and Thompson, 1987).

Continental-rift–seated lakes are generally located along the rift axis and are bounded by sharp topography gradients perpendicular to the axis. Therefore, climate change or tectonic repositioning of the sill may lead to rapid and high-amplitude fluctuations that can induce contrasting hydrodynamic regimes. Topography-driven flow is the principal mechanism driving groundwater from the surrounding aquifers toward the rift drainage basin, forming groundwater-effluent lakes (Person and Garven, 1992; Garven, 1995). Large salinity gradients between saline lakes and groundwater in their margins produce density-driven flow, thereby forming groundwater-influent lakes. The magnitude and impact of this effect in sedimentary basins is relatively large (Evans and Nunn, 1989; Deming and Nunn, 1991; Senger, 1993; Sanford and Wood, 1995; Gupta and Bair, 1997; Lahm et al., 1998) and is especially important in continental rifts (Person and Garven, 1994; Stanislavsky and Gvirtzman, 1999).

Several previous studies have postulated, based on the chemistry of groundwater, that solutes discharging into the Sea of Galilee were derived from Lake Lisan, which covered a large area in the Dead Sea rift during the upper Pleistocene. However, these studies lacked a quantitative mechanism to allow such a process. In this study, we postulate that dense saline water percolated from Lake Lisan into the subsurface of the Kinarot Basin (Fig. 1) during a highstand phase of the lake.'' Since the recession of Lake Lisan and the instantaneous formation of the fresh-water Sea of Galilee in the Kinarot basin, the previously intruded brine has been flushed backward toward the lake. However, this hypothesis, which suggests a transition from a groundwater-influent lake to a groundwater-effluent lake, is questionable for the following reasons. (1) A groundwater-influent lake requires that the equivalent fresh-water head in the lake be higher than in the surrounding groundwater system, whereas the groundwater system between the Galilee Mountains and the Golan Heights is dominated by topography-driven flow toward the lake. (2) No evaporites of the Lisan Formation, which would have been deposited from Lake Lisan, are found around the lake. (3) The differences in groundwater chemistry between clusters of springs around the lake suggest a nonuniform source for the emerging solutes (Kolodny et al., 1999).

In the present study, we compile evidence that supports the occupation of the Kinarot basin by saline Lake Lisan during the Pleistocene. A numerical model of transient fluid flow, heat, and solute transport examines the conditions re-

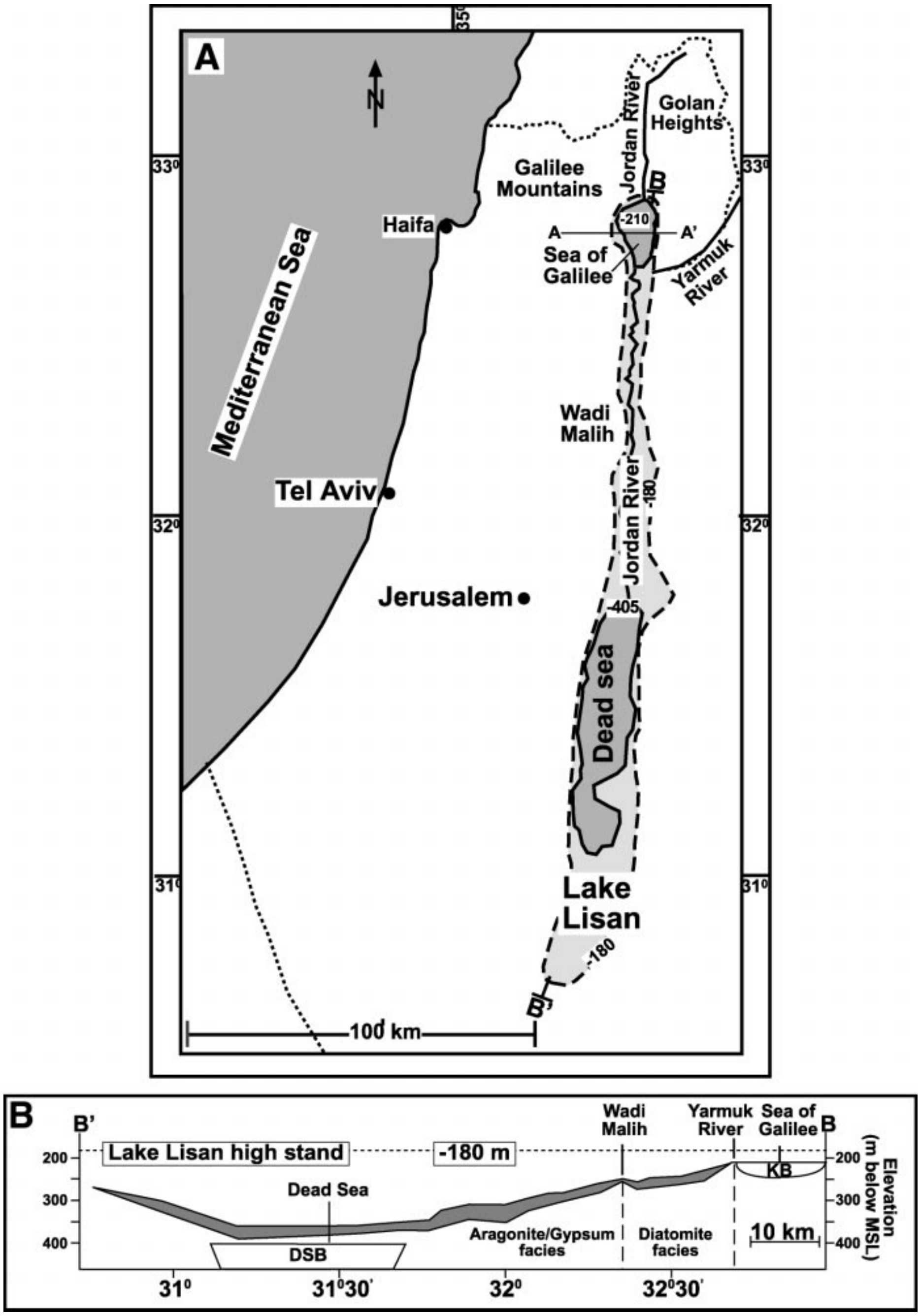

Figure 1. (A) Location map showing prior and current lakes along the Dead Sea rift. The patterned area bounded with the dashed line denotes the maximum aerial coverage of the late Pleistocene Lake Lisan during its highstand 180 m below mean sea level (m b.s.l.) (after Begin et al., 1974). The numbers along the lake margins are their stands (in m b.s.l.). Line A–A′ is a cross section presented in Figure 2, and line B–B′ is shown below. (B) A north-south cross section through Lake Lisan. The patterns are areas with outcrops of Lisan Formation sediments. The dashed vertical lines are the boundaries between the subbasins. DSB—Dead Sea basin, KB—Kinarot basin (after Begin et al., 1974).

quired for the hypothesis that suggests Lake Lisan water percolation. A parametric analysis is carried out to explore the sensitivity of the hydrodynamics to variation of the various parameters in different parts of the cross section. Simulations show the complex interplay between topography-driven and density-driven flows caused by changing lake salinity. These illustrate the transient response of groundwater systems and lakes in continental rifts to regional and global climate changes.

PALEOHYDROLOGY

The Dead Sea rift (transform) is the deepest terrestrial site on Earth. This rift extends more than 1000 km and separates the Sinai subplate from the Arabian plate. Left-lateral displacement of 105 km has occurred along the transform since its formation ca. 20 Ma (Garfunkel, 1981). The rift serves as a base level to which fresh surface water and groundwater drain from both the east and the west. An invasion of the Mediterranean Sea through the Yizrael Valley during the late Miocene formed an elongated seawater arm that covered the subsiding rift valley (Zak, 1967). After extensive evaporation in the ‘‘Sedom Lagoon’’ and dolomitization in the aquifers, a residual calcium chloride brine was formed in the subsiding rift (Starinsky, 1974).

A subsequent series of north-south–trending axial lakes was formed in the rift during its evolution. During the late Pliocene and the Pleistocene, lakes that covered parts of the northern Dead Sea rift adjacent to the Kinarot basin contained fresh water as inferred from their sedimentary deposits (Rosenthal et al., 1989; Shaliv, 1991). Saline Lake Lisan occupied an area from south of the Dead Sea to the Kinarot basin from 70 to 17 ka in the late Pleistocene (Fig. 1; Kaufman et al., 1992; Schramm et al., 2000). The lake levels fluctuated during this time between a minimum of 500 m below mean sea level (b.s.l.) and a maximum of 180 m b.s.l. (Yechieli et al., 1993). The lake probably attained its last highstand of 180 m b.s.l. approximately 26 ka; this highstand lasted for a period of no more than a few thousand years (Bartov, 1999). During its highstand, the lake covered several subbasins (Fig. 1B) with an area of approximately 2000 km^2 and drained a basin of 40 000 km^2. Lisan Formation sediments, which were deposited from Lake Lisan, indicate that the lake had spatial and temporal salinity variations (Begin et al., 1974). In the southern part, the sedimentary sequence consists mainly of evaporites, and salinity ranged between 100 and 300 g/L of total dissolved solids (TDS) (Katz et al., 1977; Katz and Kolodny, 1989; Stein et al., 1997). In the Kinarot basin, sediments of the Lisan Formation are extremely different, and the sequence is mainly diatomite (Fig. 1B; Begin et al., 1974) and sparse evaporites (Braun, 1991). The subbasins were connected through narrow and shallow passages in the areas of the Wadi-Malih and Yarmuk River deltas (Fig. 1B). During the

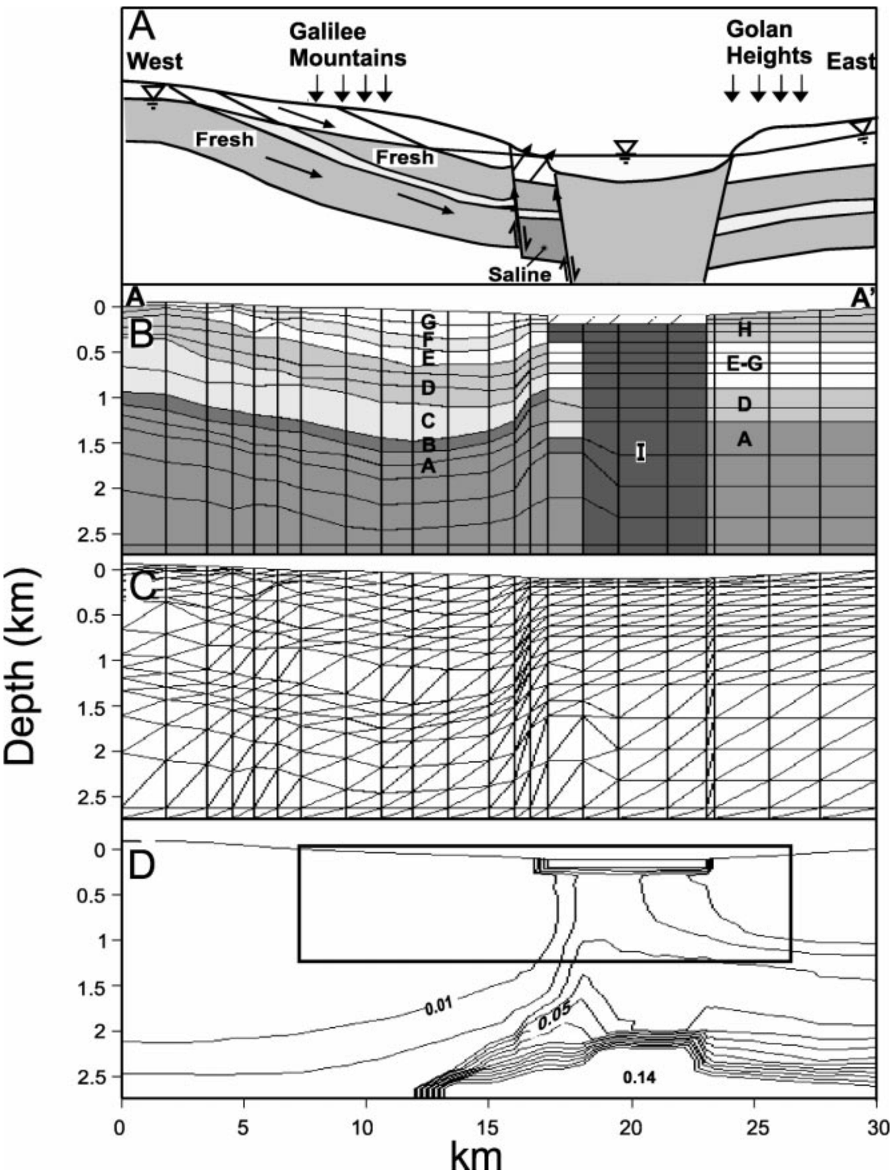

Figure 2. (A) Schematic hydrogeologic cross section from the Galilee Mountains through the Sea of Galilee to the Golan Heights. (B) Hydrostratigraphic cross section (line A–A′, Fig. 1). Formations were grouped into units based on their hydraulic properties. Unit parameters are listed in Table 1. (C) Finite element mesh. (D) Initial salt distribution established using a virtual 500 000 yr simulation. Contours are solute concentration (mass fraction). Contour intervals are 0.01 g/g. The box surrounding the lake is the area presented in Figures 3–8.

highstand, flow was partially obstructed by sills.

Lake Lisan was density stratified throughout most of its existence (Stein et al., 1997). The following evidence indicates that the hypolimnion in the northern subbasin was also saline during the lake's highstand phase. (1) Sediment cores recovered from the bottom of the Sea of Galilee contain abundant brackish- and hypersaline-water diatom species (Erlich, 1985). (2) Sediment-hosted brine was detected at a depth of 10–15 m below most of the lake bottom in a recent geophysical study (Hurwitz et al., 1999). (3) Aragonite deposits, dated to 26 $\pm$ 3 ka, were found 20 km south of the Sea of Galilee (Braun, 1991). (4) The saline component of groundwater emerging in springs contains trace amounts of ^{14}C, implying that it was discharged into the subsurface later than ca. 30 ka (Bergelson et al., 1999). (5) The sills separating the subbasins of Lake Lisan did not prevent flow from the Dead Sea basin northward as indicated by the lack of post-Lisan deformation (Rotstein et al., 1992; Ben-Gai and Reznikov, 1997). However, salinity in the Kinarot basin was lower than 100 mg/L as indicated both by the lack of evaporites and by oxygen isotopes (Gat et al., 1969; Bergelson et al., 1999).

Two separated lakes formed upon the decline of the high Lake Lisan levels. These are the hypersaline terminal Dead Sea in the south (~340 g/L TDS) and the flow-through Sea of Galilee in the north (~0.45 g/L TDS). Lake elevations are 410 m b.s.l. and 210 m b.s.l., respectively (Fig. 1B). This event probably occurred between 20 and 18 ka (Horowitz, 1979; Nadel et al., 1995). It is suggested that the transition from a saline lake to a fresh-water lake did not take more than a few tens of years.

The Sea of Galilee (Fig. 1) has a surface area of approximately 170 km^2 and a maximum water depth of 47 m (Ben-Avraham et al., 1990). Currently, saline groundwater enters the freshwater lake by advection through the faulted rift margins (Goldshmidt et al., 1967; Gvirtzman et al., 1997a; Rimmer et al., 1999) and by diffusion through the bottom sediments (Stiller, 1994; Stiller and Nissenbaum, 1996; Hurwitz et al., 2000). Onshore and offshore springs have chloride concentrations that range from 3000 to 22 000 mg/L (Mazor and Mero, 1969; Bergelson et al., 1999; Kolodny et al., 1999). This range represents variable mixtures between a saline component emerging from depth and meteoric water emerging from shallow aquifers. The maximum TDS and most ionic ratios are similar in the groundwater on both sides of the lake, with the exception of the calcium to magnesium ratio. In sources on the eastern side, the ratio is between 0.2 and 0.6, whereas, on the western side, it is above unity, 1.5 to 3.1 (Bergelson et al., 1999; Kolodny et al., 1999).

HYDROLOGIC MODELING

Previous numerical modeling of the Kinarot basin (Gvirtzman et al., 1997a, 1997b) used steady-state hydrodynamics to simulate fluid flow and heat transport. In this study, we incorporate several modifications for a transient system. These assess the effects of fluctuating lake stands on fluid migration and salinity redistribution in the vicinity of the lake.

The model geologic cross section is 2.7 km deep by 30 km long, extending from the water divide in the Galilee Mountains in the west, through the Sea of Galilee, to the water divide in the Golan Heights in the east (Fig. 2). The geologic data were compiled from the sources listed in Gvirtzman et al. (1997a, 1997b). The integrated stratigraphic sequence was divided

TABLE 1. PHYSICAL PARAMETERS OF HYDROSTRATIGRAPHIC UNITS

Symbol	Age	Group or formation	Lithology	Porosity	K_h* (m/yr)	K_v† (m/yr)	λ§ (W/°C/m)
I	Neogene–Quaternary	Dead Sea–Tiberias Groups	Clastics	0.10	0.5	0.5	2.5
H	Eocene	Zora Formation	Chalk	0.10	40	1.3	3
G	Cretaceous	Sakhnin Formation (Upper JGA)	Dolomite, Limestone	0.15	200	7.0	3
F	Cretaceous	Deir-Hanna Formation (Middle JGA)	Chalk, Limestone	0.10	80	2.7	3
E	Cretaceous	Kamon Formation (Lower JGA)	Dolomite, Limestone	0.15	200	7.0	3
D	Cretaceous	Upper Kurnub Group (Upper KGA)	Marl, Limestone	0.10	40	4.0	3
C	Cretaceous	Lower Kurnub Group (Lower KGA)	Limestone, Sandstone	0.15	80	2.7	3
B	Cretaceous	Tayasir Volcanics	Basalt	0.10	5	1.7	3
A	Jurassic	Arad Group	Limestone	0.10	40	1.3	3

*K_h is horizontal hydraulic conductivity.
†K_v is vertical hydraulic conductivity.
§λ is thermal conductivity.

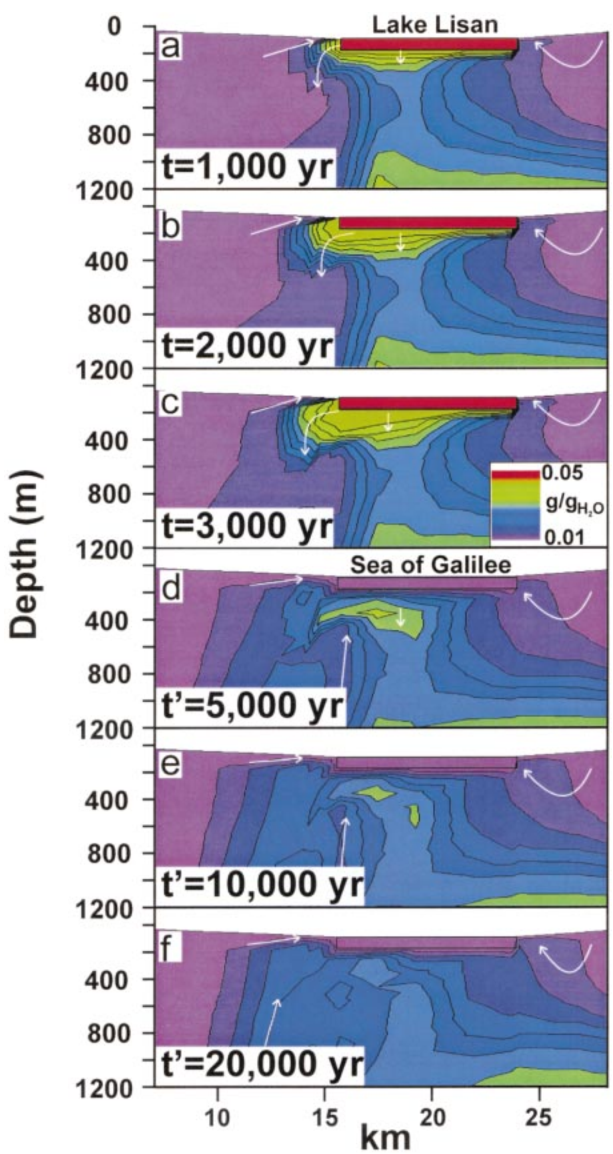

Figure 3. **Simulation results depicting salinity distribution and flow direction during the highstand phase of former saline Lake Lisan (lake salinity 0.05 g/g) and during the subsequent stages when the basin was occupied with the fresh-water Sea of Galilee (lake salinity 0.005 g/g). Salt distribution after (A) 1000 yr, (B) 2000 yr, (C) 3000 yr, during which times Lake Lisan covered the Kinarot basin. Salt distribution after (D) 5000 yr, (E) 10 000 yr, (F) 20 000 yr, during which times the Sea of Galilee covered the Kinarot basin.**

into nine hydrostratigraphic units according to their hydraulic properties (Table 1 and Fig. 2). These properties were obtained from measurements of core samples, pumping tests (Bein, 1967; Nativ and Menashe, 1991), and global compilation of various lithologies (Freeze and Cherry, 1979, p. 29). The units in the margins range from the Upper Jurassic Arad Group at the bottom to the Eocene Zora Formation exposed in the Golan Heights. The change in lithology between the two sides of the rift results from the 105 km, left-lateral shift along the Dead Sea transform (Freund et al., 1970). The sequence within the rift consists of more than 4 km of Miocene to Quaternary fluvial and lacustrine clastic and carbonate rocks, basalts, and evaporites (Marcus and Slager, 1985).

Simulations were conducted using the finite-element code JHU2D (Garven, 1989), which allows simulations of a transient flow pattern for variable-density groundwater. At each time step, a new salt concentration was calculated in each node using the flow velocity known from the previous time step and the salt-concentration gradient calculated using linear interpolation between adjacent nodes. For the grid size used, longitudinal and transverse dispersivities of 200 m and 20 m, respectively, were applied for all hydrostratigraphic units to reflect the large scale of flow (Gelhar et al., 1992).

The hydrostratigraphic cross section forms a grid of 16 nodal rows by 23 nodal columns forming 330 elements, each defined with hydraulic and thermal parameters (Fig. 2, A and B, and Table 1). Hydraulic boundary conditions were atmospheric pressure at the groundwater table and no-flow conditions at the base and on the western and eastern sides. A constant temperature of 20 °C was assigned to the water table, and a steady heat flux of 45 mW/m² was designated at the bottom (Eckstein and Simmons, 1978). Thermally insulated boundaries were assumed on both sides of the cross section. Boundary conditions of water salinity were defined as zero on the bottom and side boundaries and assumed to be 0.005 g/g at the top boundary, where rainwater infiltrates through the land surface. The exception to the latter is the area representing the lake itself, where its salinity was defined by the salinities of Lake Lisan and the Sea of Galilee. The nodes representing the area around the salt layers at depth were assigned a constant salinity of 0.14 g/g and held constant throughout the simulation. All other nodes in the cross section were allowed to change with time. Prior to the occupation of the Kinarot basin by Lake Lisan, a steady-state distribution in the cross section with a maximum concentration of 0.025 g/g was established using a simulation representing 500 k.y. of transient groundwater flow, heat transport, and solute redistribution (Fig. 2C).

Four simplifying assumptions were incorporated in the proposed model to obtain a better insight of the parameters that control basin hydrodynamics. (1) The three-dimensional, nearly radial flow field surrounding the Sea of Galilee was neglected. The two-dimensional cross section (Fig. 2A) is parallel to the subsurface hydraulic gradient in the Galilee Mountains and is probably slightly oblique to the hydraulic gradient in the Golan Heights. (2) There has been no post-Lisan tectonic activity in the Dead Sea rift, Galilee Mountains, or Golan Heights that might have significantly affected the flow field. (3) On a geologic time scale, the recession of saline Lake Lisan and its transformation into the fresh-water Sea of Galilee was instantaneous. This assumption is reasonable taking into account that the residence time of water in the Sea of Galilee is ~5 yr. (4) Compaction of the subsiding rift sediments was neglected because these transients are orders of magnitude lower in the time scale we are using (Bethke, 1985; Person and Garven, 1994; Hurwitz et al., 2000).

RESULTS

The first set of simulations depicts the processes of saline-water percolation from Lake Lisan, which covered the basin during its highstand phase (Fig. 3, A–C). Lake depth was 80 m (35 m more than the Sea of Galilee). The lake water, with an assumed salinity (for the Kinarot basin) of 0.05 g/g, progressively invaded into both the Judea Group carbonate aquifer and the Kurnub Group carbonate-sandy aquifer on the western side of the lake and into the underlying low-permeability, syn-rift sediment. No brine percolated into the Zora Formation chalk aquitard on the eastern side of lake, emphasizing the asymmetry of the flow field (Arad and Bein, 1986).

In the simulation, 3000 yr after percolation had commenced, the brine front (0.01 g/g concentration contour) was located at a lateral distance of ~3 km and at a depth of 600 m on the western side (Fig. 3C). At this stage, lake-water salinity had decreased to 0.005 g/g to represent the recession of Lake Lisan and the instantaneous formation of the fresh-water Sea of Galilee. Lake salinity was kept constant throughout the simulation, and the lake depth was reduced to 40 m. Initial salt distribution was defined as the final distribution in the first set of simulations (Fig. 3C). As a response, circulating meteoric water began flushing the saline groundwater back toward the rift margins (Fig. 3, D–F). This process caused a continuous eastward shift of the salt-water interface on the western margins of the lake. At 20 k.y. after the transition, the concentration gradient on the western side decreased due to leaching and dilution of the salt plume (Fig. 3F). At this stage, deep saline water ascended from the Judea Group carbonate aquifer and Kurnub Group carbonate-sandy aquifer, mixed with shallow meteoric water, and emerged at the surface. The concentration distribution at depth on the western side was similar to the present distribution inferred from some wells along the lake shores (Rimmer et al., 1999).

A sensitivity study was conducted to explore the effects of various parameters on displacement of saline Lake Lisan waters into the subsurface during the highstand and the subsequent backward flow toward the Sea of Galilee. Anisotropy of the hydraulic conductivity had the largest effect on simulation results (Fig. 4). In numerous basin-scale studies, assumed anisotropy ranged between 10 (e.g., Deming and Nunn, 1991; Gupta and Bair, 1997) and 100 (e.g., Garven and Freeze, 1984; Person and Garven, 1992), but these were generally poorly fixed. Therefore, the vertical hydraulic conductivity was adjusted so that—apart from the basin-fill unit (unit I, Fig. 2A and Table 1)—the anisotropy ratio in all lithostratigraphic units varied between 10 and 100. Anisotropy of 10 caused the brine front to percolate to a depth of approximately 1000 m (Fig. 4A), whereas anisotropy of 100 caused only slight percolation after 3000 yr (Fig. 4C). Upon the transition into a fresh-water lake, the brine continued percolating to depth. To achieve both significant vertical percolation and backward leaching of solutes, the optimal anisotropy was 30.

Lake-water salinity is another major parameter affecting the hydrodynamics in the basin. When initial salinity was 0.03 g/g, no percolation occurred, indicating that topography-driven flow was dominant in the system (Fig. 5A). When salinity was increased to 0.07 g/g,

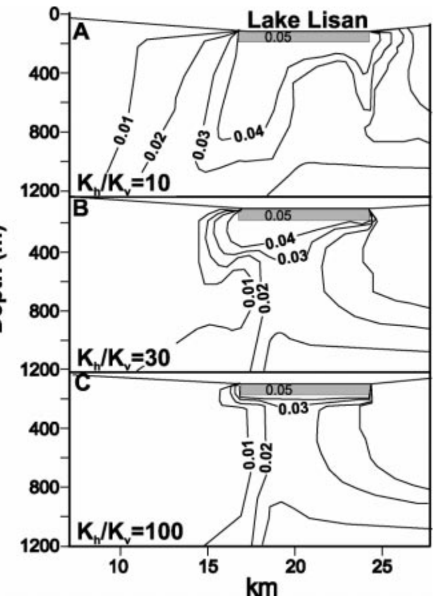

Figure 4. Effect of anisotropy on the distribution of solute concentration in the subsurface after 3000 yr, during which time the Kinarot basin was covered by Lake Lisan. The values of the horizontal hydraulic conductivity (K_h) are shown in Table 1; the anisotropy ratios of units A to H were changed to (A) 10, (B) 30, and (C) 100.

vertical percolation on the western side was extensive, and the brine front reached a depth of about 700 m after 3000 yr (Fig. 5C). An optimal simulation, where significant percolation and backward leaching of solutes occurs, was achieved with a salinity of 0.05 g/g (Fig. 5B).

We examined the effect of water-table elevation differences between the recharge areas along the water divides to the level of Lake Lisan. Coeval with the highstand period of Lake Lisan, the Mediterranean Sea level was lower than at present by 100–120 m (Shackelton, 1987). This low Mediterranean base level on the west together with the highstand on the eastern base level (Dead Sea rift) must have caused the decline of the groundwater table and an eastward shift of the water divide (Kafri and Arad, 1978), which resulted in the decrease of gradients toward the lake. However, our numerical simulations indicate that even a 30% decrease of the hydraulic gradient had a negligible effect on the rate of saline-water percolation.

A fourfold increase of the hydraulic conductivity of the Judea Group carbonate aquifer, which is the major aquifer on the western side, had a minor effect compared with the simulation presented in Figure 3C. In this situation, the brine front migrated to a similar depth (700

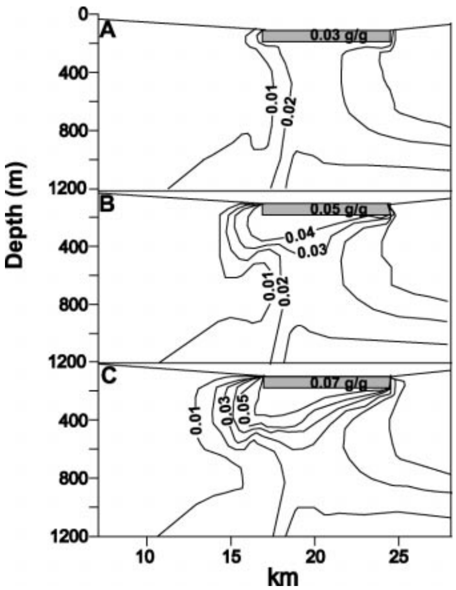

Figure 5. Effect of lake-water salinity on the distribution of solute concentration in the subsurface after 3000 yr, during which time the Kinarot basin was covered by Lake Lisan. Lake-water salinity is (A) 0.03 g/g, (B) 0.05 g/g, and (C) 0.07 g/g.

m) but a greater lateral distance of 4 km (Fig. 6A). When the hydraulic conductivity of the low permeability basin fill was decreased by an order of magnitude, the brine percolated to the same depth and lateral distance on the western side, as in Figure 3C, but there was no significant percolation into the underlying basin-fill sediments (Fig. 6B).

The effect of a 300 m fault zone on the western margin of the Kinarot basin was also examined (Fig. 7B). In this simulation, we allowed a period of 1000 yr for the brine to percolate into the subsurface. Horizontal and vertical hydraulic conductivities in the fault zone were increased to 500 m/yr. After 1000 yr, the 0.04 g/g contour shifted downward to a depth of 900 m, approximately along the trace of the fault zone with negligible lateral migration (Fig. 7B). After a 20 k.y. period during which the fresh-water (0.005 g/g) Sea of Galilee covered the basin, the concentration gradient decreased due to leaching and dilution of the salt plume (Fig. 7C), and the concentration distribution was similar to the distribution in Figure 3F.

DISCUSSION

Large climatic changes during the Pleistocene had a significant effect on the amplitude, frequency, and wave length of lake-level fluctuations. Following these climatic changes, variations of lake-water density resulted in contrasting hydrodynamic regimes between lakes and the surrounding groundwater system. In this study, we explored the parameters controlling the hydrodynamics of the Kinarot basin during two subsequent stages and determined the conditions under which the lake is either groundwater influent and dominated by density-driven flow or groundwater effluent and dominated by topography-driven flow.

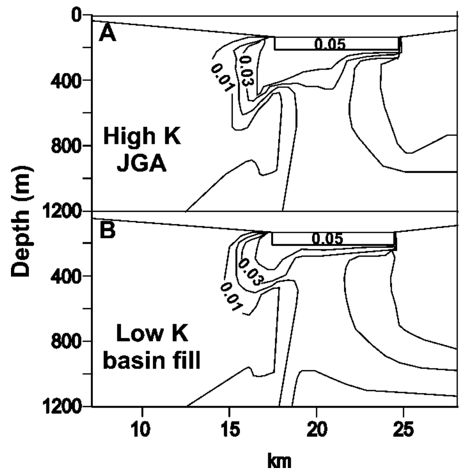

Figure 6. Effect of permeability on the distribution of solute concentration in the subsurface after 3000 yr, during which time the Kinarot basin was covered by Lake Lisan. (A) The value of the horizontal hydraulic conductivity (K_h) in the upper and lower Judea Group aquifer (JGA) was increased to 800 m/yr, and the value of the vertical hydraulic conductivity (K_v) was increased to 27 m/yr. (B) Values for the K_h and K_v of the basin-fill sediments (Dead Sea and Tiberias Groups—Unit I) were reduced to 0.05 m/yr.

The sensitivity analysis shows that a wide range of hydrodynamic conditions is compatible with the hypothesis that occupation of the Kinarot basin by Lake Lisan during the Pleistocene created an influent environment. Actually, deep percolation could occur if (1) lake-water salinity was lower than 0.03 g/g, (2) the permeability anisotropy was higher than 100, (3) the hydraulic conductivity of the basin-fill sediments (unit I, Fig. 2A) was lower than 0.05 m/yr, and (4) the lake highstand was shorter than 1000 yr. These constraints imply that subsurface seepage from Lake Lisan probably occurred. Our results show that a period of 1000–3000 yr was sufficient for Lake Lisan saline waters to percolate to depths of 700–1000 m in the western margins of the lake within the Judea Group carbonate aquifer and Kurnub Group carbonate-sandy aquifer (Fig. 4C). The

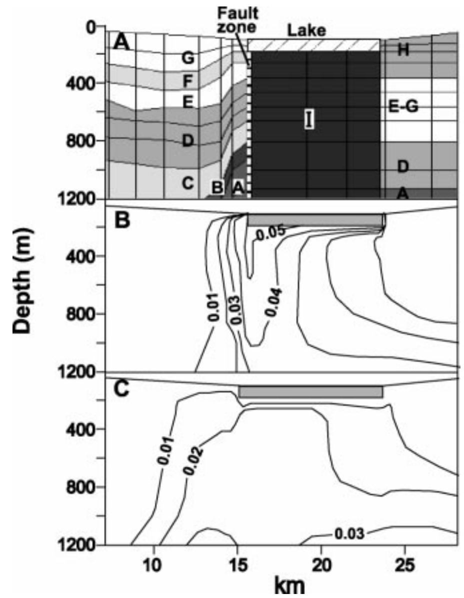

Figure 7. (A) A 300-m-wide fault zone was introduced into the cross section. The values for the horizontal and vertical hydraulic conductivities (K_h and K_v, respectively) in the fault zone are 500 m/yr. (B) The distribution of solute concentration after 1000 yr, during which time the Kinarot basin was covered by Lake Lisan. (C) The distribution of solute concentration after 20 000 yr, during which time the Kinarot basin was covered by the Sea of Galilee.

saline waters also migrated slowly within the low permeability basin-fill sediments to a depth of 200–400 m. This implies that salinity-driven flow was strong enough to overcome the topography gradients between the regional aquifers and the lake. Upon the instantaneous occupation of the Kinarot basin with a fresh-water lake, upward leaching of solutes was induced, and the system became dominated by topography-driven flow. Under a range of simulation conditions, solute distribution in the subsurface after having a fresh-water lake occupy the Kinarot basin for 20 k.y. is similar to the distribution measured in several boreholes around the lake (Rimmer et al., 1999).

Apart from the simulation where the hydraulic anisotropy was decreased to 10 (Fig. 4A), no significant percolation occurred in the eastern margin. This lack of percolation emphasizes the importance of the vertical permeability in the recharge area and the asymmetry of the groundwater flow system. Percolation of saline water into the high-permeability aquifers of the Judea Group and Kurnub Group that crop out in the Galilee Mountains was significant. In contrast, the low permeability of the Zora Formation aquitard prevented brine percolation.

Whereas there is common agreement that the origin of the emerging brine is of marine origin from the Miocene, Sedom Lagoon (Starinsky, 1974), there is an ongoing debate regarding the evolution and hydrodynamics of this brine. It was suggested that the original brine percolated into isolated pockets in the subsurface and that this brine currently is being discharged in springs along the lakeshore (Mazor and Mero, 1969; Kolodny et al. 1999). A scenario presented in this study is that the Sedom Lagoon brine (salinity of 300 g/L) either has both percolated and settled in deep parts of the section (e.g., Rosh-Pina borehole, north of the lake), or it has flowed on the surface to the Dead Sea basin, where it formed evaporites. During a highstand phase of Lake Lisan, however, the brine flowed from the Dead Sea basin to the Kinarot basin. A scenario presented in this study is that the Sedom Lagoon brine (salinity of 300 g/L) has either percolated and settled in deep parts of the section (e.g. Rosh-Pina borehole, north of the lake), or it flowed on the surface to the DSB, where it formed evaporites. Following the Sedom Lagood period, the KB was probably occupied only by fresh-water lakes (Rosenthal et al., 1989). However, during a high-stand phase of Lake Lisan brine flowed from the Dead Sea basin to the KB. Our hypothesis is supported by oxygen isotope data (Gat et al., 1969; Bergelson et al., 1999), geophysics (Hurwitz et al., 1999), hypersaline diatom species in the sediments (Erlich, 1985), and ^{14}C measured in the saline groundwater component (Bergelson et al., 1999).

Numerous studies have noticed the relatively high variability of groundwater chemistry among the different clusters of springs in the Kinarot basin (e.g., Mazor and Mero, 1969; Vengosh et al., 1994; Kolodny et al., 1999; Bergelson et al., 1999). Whereas most ionic ratios vary within less than an order of magnitude, the calcium to magnesium ionic ratio has a distinct difference between sources on both sides of the lake. On the eastern side, the ratio ranges from 0.2 to 0.6, whereas, on the western side, it is above unity and ranges from 1.5 to 3.1. The cause for this difference has not yet been elucidated. Our simulation results suggest that this difference may be explained by the asymmetric flow patterns observed in Figures 3–7. Whereas saline water percolated to depths of 600 to 1000 m on the western side, it percolated much less into the eastern margin. During the percolation period, downward flow brought relatively cold water from the surface to depths, and, therefore, a low thermal gradient formed. Upon the transition into a fresh-

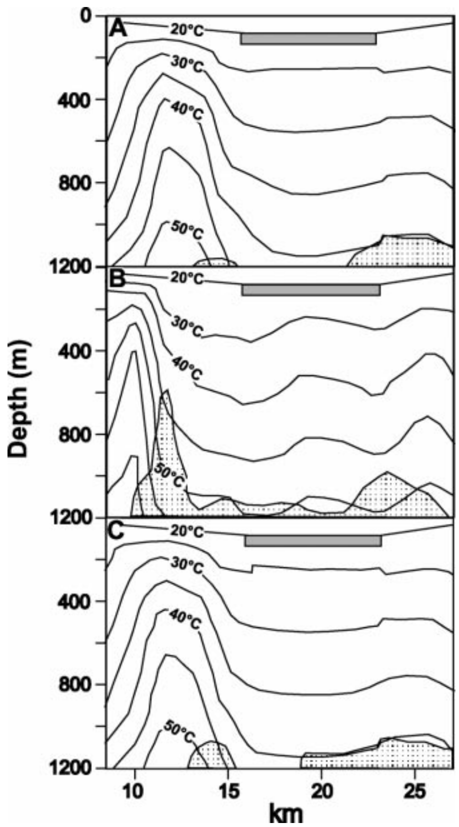

Figure 8. The temperature (contours) and the area with a salinity greater than 0.03 g/g (pattern) after 20 000 yr, during which time the Sea of Galilee covered the Kinarot basin. (A) Simulation parameters as in Table 1. (B) Anisotropy of units A–H is 10 (similar to Fig. 4A). (C) A 300 m fault zone on the western margin of the basin is introduced (Fig. 7).

water lake and accompanying upward flow of brine, a large thermal gradient formed (Fig. 8). The elevated temperatures enhanced the chemical disequilibrium of the brine with limestone aquifers of the Judea and Kurnub Groups to form epigenetic dolomite:

$$2CaCO_3 + Mg^{2+} \leftrightarrow CaMg(CO_3)_2 + Ca^{2+}.$$

As a result, the residual solution emerging in springs on the western side is enriched with calcium and depleted in magnesium. Dolomitization commenced after the transition to a groundwater-effluent lake, and this process might still be occurring because the saline waters are located at depth and have elevated temperatures. Currently, waters discharging from some of the springs along the western lakeshore originate from such depths (Rimmer et al., 1999). It should be noted also that the percolating Lake Lisan waters were already a calcium chloride brine that formed epigenetic dolomite during the Pliocene (Starinsky, 1974). Therefore, the post-Lisan stage of dolomitization is later in paragenesis. According to the hydrogeologic model calculations, the dolomitization occurs at temperatures exceeding 45 °C within the Judea Group carbonate aquifer and Kurnub Group carbonate-sandy aquifer (Fig. 7). This relatively elevated temperature helps to reduce kinetic barriers known to exist at lower temperatures (Hardie, 1987; Kaufman, 1994). We suggest that if aragonite were deposited from the northern Lake Lisan (which is not likely), it was not transformed to dolomite because of the relatively low mean temperature at the surface. Dolomitization also requires that a large volume of water be cycled through the limestone aquifer. It has been suggested that large water-level fluctuations act as a ''pump,'' inducing flow into and out of the subsurface (Whitaker and Smart, 1990).

The introduction of a fault zone into the simulated cross section had a major impact on the flow system (Fig. 7). In this particular case, we examined brine migration in the fault zone to a depth exceeding 1 km in 1000 yr (Fig. 7B). The consequent leaching of solutes was achieved within 2000 yr. We suggest that, in addition to heterogeneous subsurface lithology, a major parameter causing spatial variability of ionic ratios (e.g., sodium/chloride and bromide/chloride) is fault-zone permeability. In areas where faulting is more extensive and permeability is greater, percolation flow rate was higher, enhancing differential hydrodynamic dispersion of the different ions.

SUMMARY AND CONCLUSIONS

Applying numerical modeling, we examined and simulated various parameters affecting saline-water percolation from the Pleistocene Lake Lisan into the subsurface and the subsequent discharge into the fresh-water Sea of Galilee. Recharge of the saline water into the subsurface was induced by the high salinity of the lake, which triggered density-driven flow. After the rapid transition into a fresh-water lake, displacement of the invading Lake Lisan solutes backward toward the lake became dominated by topography-driven flow. A sensitivity analysis indicates that the major parameters affecting the hydrodynamic regime are aquifer permeability in the recharge area, anisotropy of the hydraulic conductivity, lake-water salinity, and the existence of a fault zone. Results suggest that different flow regimes on both sides of the rift have caused some major chemical differences. On the western side, saline and relatively warm groundwater circulated through the limestone aquifers of the Judea and Kurnub Groups and formed epigenetic dolomite. The residual saline groundwater, depleted in magnesium and enriched in calcium, is currently discharging through springs into the lake. On the eastern side of the lake, saline water was not recharged into the subsurface, and, therefore, Lake Lisan waters did not change their composition.

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