Hydrogeological modeling of the saline hot springs at the Sea of Galilee, Israel

Haim Gvirtzman
Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel

Grant Garven
Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, Maryland

Gdaliahu Gvirtzman
Department of Geography, Bar Ilan University, Ramat Gan, Israel

Abstract. Meteoric fresh groundwater from shallow aquifers and hot brines from deep aquifers mix while emerging from several springs along the western coast of the Sea of Galilee, a freshwater lake located within the Dead Sea Rift Valley, Israel. After the rainy season, when elevations of the groundwater table rise in the regional aquifers and discharge rates of springs increase, solute concentrations decrease at Tabha springs but, surprisingly, increase at Fulya springs, apparently suggesting two different salinization mechanisms. Two detailed geologic cross sections were constructed, one across the rift valley at Tabha and a second at Fulya, each about 6 km deep and 70 km long. The hydrodynamics in these cross sections were analyzed using a two-dimensional finite element code that solves the coupled variable-density groundwater flow and heat transfer equations. Numerical simulations indicate that a topography-driven flow model explains both spring systems, and the opposite salinity behavior results from the different hydrogeological configurations of the two subsurface drainage basins. At Fulya, both aquifers, the shallow one and the deeper one, are partially phreatic, whereas at Tabha, the deeper aquifer is totally confined. The response of springs to changes in elevation of groundwater table were simulated, reproducing field observations. This analysis has implications for the management scheme for the lake and its surrounding aquifers.

Introduction

The Dead Sea Rift Valley (Figure 1) is the deepest terrestrial location on Earth and includes 5000 km² with elevations below sea level. This depression serves as a base level to which fresh surface waters drain. Two lakes are located within the rift: the freshwater Sea of Galilee (known in Israel as Lake Kinneret) and the saline Dead Sea. Surface water elevations are ~210 m and about ~400 m (below mean sea level), respectively. Lake Kinneret is fresh because its annual fresh inflow and outflow water volumes are equal to about 20% of its total capacity. On the other hand, the Dead Sea is saline because it is a closed basin lake, with no outflow except evaporation.

The rift valley depression is also the base level to which groundwater discharges through springs. Unlike surface water streams, groundwater is a mixture of fresh and saline sources. Based on geochemical analyses of waters collected at wells and springs at the lake vicinity, it appears that the fresh component drains laterally from regional shallow aquifers at both sides of the valley, while much of the saline component originates from hot brines situated in deeper aquifers [Mazor, 1968; Starinsky, 1974; Fleischer et al., 1977; Starinsky et al., 1979].

Saline hot springs emerge mainly at three locations along the western coast of Lake Kinneret: Tabha, Fulya, and Tiberias (Figure 1). Their chloride concentrations range from about 1000 to 18,000 mg/L, depending on the season and location. These saline springs pose a major problem to Israel's water system, because Lake Kinneret supplies about 500 million m³ of water per year, or about 30% of the country's annual consumption [Tahal, 1988]. The onshore and offshore saline water springs supply an average of 160 million kg of chloride per year to the lake [Simon and Karni, 1976]. However, the saline water aqueduct, constructed in 1967, diverts part of the coastal spring water (17 million m³ of water containing about 55 million kg of chloride per year) away from the lake, which has resulted in reducing the chloride concentration in the lake from its original value of 400 to about 220 mg/L. From a practical point of view, the lake’s salinity is too high for irrigating some crops, and when pumped lake water is transferred into the coastal plain aquifer for seasonal storage, it increases the aquifer’s salinity.

In order to further reduce the lake’s salinity, it is of critical importance to better understand the salinization mechanism; that is, what is the driving force for the ascent of the hot brines from aquifers, over 2 km deep, to the surface? Moreover, to prevent inflow of brines into the lake, the National Water Company decided that the water elevation in the lake should not be allowed to drop below ~213 m. If one could safely lower...
During the past 45 years, three conceptual models have been proposed (Figure 2) to address the salinization mechanism in Lake Kinneret: (1) the “self-potential brine” model, (2) the “stagnant brine,” or “leaching,” model, and (3) the “sea water intrusion” model (all reviewed by Simon and Mero [1992]). The self-potential brine model [Mero and Mandel, 1963; Mazor and Mero, 1969] hypothesizes that deep brines are overpressured either by (1) compaction of sediments accumulated in the subsiding rift, (2) tectonic movement along the rift faults, (3) a geothermal source, or (4) pressure buildup due to degradation of organic matter. According to the stagnant brine model [Goldsmith et al., 1967], the deep brines are not overpressured, but circulating meteoric groundwater, recharged at eastern Galilee, leaches brines from deeper horizons. The third model [Kafri and Arad, 1979] proposes current Mediterranean Sea water intrusion, which displaces the deep saline groundwater towards the lake because of the 210-m hydraulic head difference between the Sea’s and the lake’s surface levels (although a local hydrological divide exists in upper aquifers). None of these models has been quantitatively evaluated using flow and transport numerical simulations along actual geologic cross sections.

On the basis of these conceptual models, different inferences have been made regarding the effects on lake salinity of pumping groundwater in eastern Galilee [Mercado and Mero, 1984]. According to the self-potential brine model (Figure 2a), pumping fresh groundwater from shallow aquifers will reduce the discharge of the freshwater component through the springs, whereas the discharge of the saline-water component would not be changed. Thus spring salinity will increase, which obviously is not desirable. In contrast, according to the leaching model (Figure 2b), pumping groundwater at the eastern Galilee would be recommended as it would reduce the volume of circulating fresh water which leaches brines into the lake. From this practical point of view, the third model (Figure 2c) is not different from the first one and we will only refer to two major conceptual models.

The two competing models were based on observations regarding the relationships between discharge and salinity at the springs [Kahanovich and Mero, 1973; Mercado and Mero, 1984; Simon and Mero, 1992]. After the rainy season, when the elevations of the groundwater table rise in the regional aquifers and discharge rates of springs increase, solute concentrations decrease at Tabha, increase at Fulya, and do not vary at Tiberias. Figure 3 exemplifies the typical inverse correlation between discharge and salinity at Tabha at the artesian borehole, Kinneret 7. During the 1991–1992 winter, as the discharge rate increased from 0.13 to 0.20 m$^3$/s, chloride concentrations decreased from 1200 to 100 mg/L. In contrast, Figure 4 exemplifies the typical direct correlation at Fulya at spring 6. As discharge rate increased from 0.05 to 0.15 m$^3$/s, chloride concentrations increased from 500 to 1500 mg/L. These direct and inverse relationships are expressed not only at the seasonal scale but also at several times during short-term fluctuations due to specific storms. Figure 5 shows that at two typical springs of the Tiberias group, chloride concentration stays almost constant through time; it ranges between 17,500 and 19,000 mg/L. Unfortunately, discharge rates were not measured in these springs.

It should be noted that without artificial pumping, which took place during the 1950s and 1960s, these three types of discharge-versus-salinity relationships were clearly detected [Kahanovich and Mero, 1973]. During the 1970s and 1980s, as groundwater pumping increased at eastern Galilee, these relationships were slightly smoothed. Nevertheless, the data presented in Figures 3–5, from hydrological years 1991–1992 and 1992–1993 (J. Geifman, personal communication, 1995), reflect mostly natural characteristics, as shown by the winter of 1991–1992. This was the rainiest year of the last several decades (250% more rainfall than in an average year), and elevations of the groundwater table increased by tens of meters in several wells at the eastern Galilee, and thus pumping caused a negligible influence on the groundwater table.

The self-potential brine model (Figure 2a) was developed on the basis of observations at Tabha (Figure 3). It was suggested that the deep overpressured saline groundwater source is diluted by the fresh one after the rainy season, when elevations of the groundwater table increase in the regional aquifers. On the other hand, observations at Fulya (Figure 4) suggest that brine discharge is enhanced by increasing the elevation of groundwater table after the rainy season and can be better explained by the leaching model (Figure 2b). It appears that each of the two conceptual models explains the behavior of one
Figure 2. Schematic E-W cross sections illustrating the conceptual models suggested for the emergence of the saline hot water through springs at Lake Kinneret: (a) the “self-potential brine” model, according to which deep brines are over pressured due to compaction of sediments and/or tectonic movements; (b) the “stagnant brine,” or “leaching,” model, according to which brines are pushed by the circulating meteoric groundwater; and (c) the “seawater intrusion” model, according to which the driving force is the 210-m difference between the Mediterranean Sea and the lake. Figures 2a–2c were prepared on the basis of concepts suggested by Mazor and Mero [1969], Goldshmidt et al. [1967], and Kafri and Arad [1979], respectively.
Thetwogroupsofspringsare7km apart, both contain mix-
tures (at different ratios) of almost the same brine and mete-
oricsources, and the geologic structure at both places is very
similar; thus it is unreasonable to arguethat they belong to two
independent and unconnected hydrological systems. Therefore
themain objective of this studyisto construct an integrated
model to explain the entire hydrological system, including both
groups of springs.

Hydrogeologic Setting

Two detailed geological cross sections (A-A’ and B-B’; Figure 1)
were prepared. These traverse the Galilee Moun-
tains, Lake Kinneret, and the Golan Heights: one at the Tabha
springs (Figure 6) and the other at the Fulya springs (Figure 7).
Each section is 6 km deep by 70 km long. A list of lithostrati-
graphic units, ranging in age from Triassic to Quaternary, is
appended (Table 1). The cross sections were prepared using
subsurface data collected from 20 deep boreholes and from
several hundred kilometers of seismic lines carried out during
oil exploration in northern Israel. A map of the top of the
Judea group [Klang and Gvirtzman, 1987; Oil Exploration In-
vestments Ltd., unpublished reports] served as a reference
structural surface, and isopach maps of various pre-Judea
stratigraphic intervals were used to construct these cross sec-
tions. Lateral facies changes, such as the transition within the
Kurnub group (units 4b and 4c); wedge-outs, such as the Asher
volcanics (unit 1a); and Rosh Pina Formation (unit 2a) were
included. The asymmetry of the stratigraphic sequence on the
eastern and western sides of the rift valley is related to the
100-km shift along the transfor m [Garfunkel, 1981; Freund et
al., 1970]. Geological maps and surface columnar sections in
northern Israel [Golani, 1961; Zaltzman, 1964; Eliezi, 1965;
Michelson et al., 1987; Shaliv, 1991] were integrated with the
subsurface information while preparing these cross sections.
The stratigraphy within the rift valley is based on work by
Marcus and Slager [1985]. The internal structure of the down-
faulted blocks in the rift valley were illustrated in accordance
with Rotstein and Barton [1989] and Rotstein et al. [1992]. It
should be emphasized that the integrated stratigraphic se-
quence is grouped and divided according to their estimated
hydraulic properties into various hydrostratigraphic units (Ta-
ble 1; Figures 6 and 7).

Mesozoic to Tertiary sediments crop out in the highlands on
both sides of the rift valley and constitute the main recharge
area for the major aquifers. Three regional aquifers contain
most of the groundwater discharge to the rift base level: (1) the
600-m-thick Cretaceous Judea group (unit 5b in Table 1), of
predominantly limestone and dolomite; (2) the 400-m-thick
Lower Cretaceous Kurnub group (units 4b and 4c in Table 1),
of mainly continental sandstones; and (3) the 2500-m-thick
Jurassic Arad group (units 1b, 2b, and 3 in Table 1), of mainly
carbonates.

The subsiding rift valley is capped by a Miocene-Pleistocene
sequence, at least 4 km thick, consisting of evaporites, alluvial
deposits, basalt, and a few intrusions of gabbro [Marcus and
Slager, 1985]; these rocks are mainly aquitards. This sequence
confines flow through the regional aquifers, forming two inde-
pendent aquifer systems at both sides of the rift. Indeed,
groundwater systems at the sides of the graben behave differ-
ently [Arad and Bein, 1986]. On the western margin of the

Figure 3. Chloride concentration and discharge versus time
at an artesian borehole, Kinneret 7, at Tabha during 1991–
1992 and 1992–1993. The general pattern is an inverse corre-
lation between discharge and salinity. Discharge rates include
±0.01 m³/s measurement errors. Data collected by the Mekorot
Company (J. Geifman, personal communication, 1995).

Figure 5. Chloride concentration versus time at two of Tibe-
rias saline hot springs during the years 1991–1992 and 1992–
1993. Data collected by the Mekorot Company (J. Geifman,
personal communication, 1995).

Figure 4. Chloride concentration and discharge versus time
general pattern is a direct correlation between discharge and
salinity. Discharge rates include ±0.02 m³/s measurement er-
rors. Data collected by the Mekorot Company (J. Geifman,
personal communication, 1995).
graben, some downfaulted blocks expose the Judea aquifer (unit 5b; Figure 6) along the margins of Lake Kinneret, channeling the main discharge of the system. Moreover, since continuity between aquifer units exist through downfaulted blocks (units 4c, 11, and 5b; Figure 6 and 7), groundwater from deep aquifers flow upwards and emerge as springs. On the other hand, on the eastern margin of the graben, the low-permeability chalk, salt, and marl deposits (units 6, 8, 10, and 12; Figures 6 and 7) block discharge from the regional aquifers to the outlets.

### Hydrogeologic Modeling

Large-scale hydrogeologic models provide a useful approach for studying the nature of fluid migration in sedimentary basins, especially where the complexities of heterogeneity, struc-
The evolution of hydrothermal ore deposits has previously taken place in many sedimentary basins around the world. OilGen, a personal IRIS 4D/35 workstation, has been used for numerical codes here. It allows for the representation of flow lines and equivalent freshwater head and entrapment of the mineral deposit. Published examples where the OilGen code has previously been applied include Haszeldine and McKeown, 1995; petroleum migration of the Jurassic, 1989; and hazardous waste disposal. Other applications include Garven, 1984; the estimation of the Jurassic, 1992; and the simulation of the Jurassic, 1995.

### Table 1. Lithostratigraphic Units

<table>
<thead>
<tr>
<th>Age</th>
<th>Symbol</th>
<th>Group</th>
<th>Formations</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quaternary</td>
<td>12</td>
<td>Dead Sea</td>
<td>Lisan, “fill”</td>
<td>marl</td>
</tr>
<tr>
<td>Pliocene</td>
<td>11</td>
<td>Dead Sea</td>
<td>Upper Basalt</td>
<td>basalt</td>
</tr>
<tr>
<td>Pliocene</td>
<td>10</td>
<td>Dead Sea</td>
<td>unnamed</td>
<td>marl</td>
</tr>
<tr>
<td>Pliocene</td>
<td>9</td>
<td>Saqiye</td>
<td>Yafod</td>
<td>marl</td>
</tr>
<tr>
<td>Pliocene</td>
<td>8</td>
<td>Dead Sea</td>
<td>Sedom</td>
<td>salt, gabbro</td>
</tr>
<tr>
<td>Miocene</td>
<td>7</td>
<td>Tiberias</td>
<td>Herods, lower Basalt</td>
<td>marl, sandstone, basalt</td>
</tr>
<tr>
<td>Eocene</td>
<td>6b</td>
<td>Avdat</td>
<td>Ma’alul, Bar Kohva</td>
<td>chalk, limestone</td>
</tr>
<tr>
<td>Upper Cretaceous</td>
<td>6a</td>
<td>Mt. Scopus</td>
<td>Ein Zetim, Ghareb, Taqiye</td>
<td>chalk, marl</td>
</tr>
<tr>
<td>Upper Cretaceous</td>
<td>6</td>
<td>Shefela</td>
<td>Mt. Scopus, Adulam, Maresha</td>
<td>chalk, marl</td>
</tr>
<tr>
<td>Eocene</td>
<td></td>
<td>Albian-Turonian</td>
<td>5b Judea, upper</td>
<td>Yagur/Kamon, Dir Hana, Sakhnin, Bina dolomite, limestone, marl</td>
</tr>
<tr>
<td>Albian</td>
<td>5a</td>
<td>Judea, lower</td>
<td>Yakhini, Hidra, Asfuri, Zalmon/Rama marl, limestone</td>
<td></td>
</tr>
<tr>
<td>Lower Cretaceous</td>
<td>4c</td>
<td>Kurnub, E</td>
<td>Hatira, Nebi Said, En el Assad sandstone, limestone, marl</td>
<td></td>
</tr>
<tr>
<td>Lower Cretaceous</td>
<td>4b</td>
<td>Kurnub, W</td>
<td>Helez, Telamim, Yavne</td>
<td>limestone, sandstone</td>
</tr>
<tr>
<td>Lower Cretaceous</td>
<td>4a</td>
<td>Kurnub</td>
<td>Tayassir Volcanics</td>
<td>basalt, pyroclastics</td>
</tr>
<tr>
<td>Upper Jurassic</td>
<td>3</td>
<td>Arad</td>
<td>Zohar, Halutza</td>
<td>limestone</td>
</tr>
<tr>
<td>Middle Jurassic</td>
<td>2b</td>
<td>Arad</td>
<td>Sederot</td>
<td>limestone, dolomite, marl</td>
</tr>
<tr>
<td>Middle Jurassic</td>
<td>2a</td>
<td>Arad</td>
<td>Rosh Pina</td>
<td>marl</td>
</tr>
<tr>
<td>Lower Jurassic</td>
<td>1b</td>
<td>Arad</td>
<td>Nirim</td>
<td>dolomite, limestone</td>
</tr>
<tr>
<td>Lower Jurassic</td>
<td>1a</td>
<td>Arad</td>
<td>Asher volcanics</td>
<td>basalt, pyroclastics</td>
</tr>
</tbody>
</table>

See Figures 6 and 7.

Conservation of fluid mass is defined by

\[ \nabla \cdot (\rho_f \vec{q}) = 0 \]  

(1)

where \( \rho_f \) is the fluid density, and \( \vec{q} \) is the specific flux, which is defined by Darcy’s law

\[ \vec{q} = \vec{k} \rho_f g \left( \nabla h + \rho_0 \nabla z \right) \quad (2) \]

where \( \vec{k} \) is the intrinsic permeability tensor, \( \rho_0 \) is a reference fluid density, \( \rho_0 \) is the dynamic viscosity, \( g \) is the acceleration due to gravity, \( h \) is equivalent fresh water hydraulic head, and \( z \) is elevation above datum. The relative fluid density, \( \rho_r \), is defined by

\[ \rho_r = (\rho_f - \rho_0)/\rho_0 \quad (3) \]

Conservation of thermal energy, including both conduction and convection processes, under steady state conditions can be expressed as

\[ \nabla \cdot (\hat{\lambda} \nabla T) - \rho_r c_r \vec{q} \cdot \nabla T = 0 \quad (4) \]

where \( \hat{\lambda} \) is the effective thermal conduction-dispersion tensor of the porous material, \( T \) is temperature, and \( c_r \) is the specific heat capacity of the fluid. Regional flow of variable-density groundwater is best simulated using the stream function \( \Psi(x, z) \) representation of flow lines and equivalent freshwater head [Bear, 1972].

Computations were conducted on a Silicon Graphics personal IRIS 4D/35 workstation. The numerical code used here is OilGen [Garven, 1989], which has been successfully applied previously to many sedimentary basins around the world. The evolution of hydrothermal ore deposits [Garven et al., 1993; Raffensperger and Garven, 1994], petroleum migration and entrapment [Person and Garven, 1992], and hazardous waste burial [Haszeldine and McKeown, 1995] are some published examples where the OilGen code has previously been applied. This two-dimensional code uses the finite element method to solve the steady state fluid and heat flow.

At the initial stage of the simulation the steady state hydraulic head distribution is computed first by assuming there are no salinity or temperature gradients. Then, Darcy velocities are computed for each element in the mesh. The steady state heat equation is solved next to find the temperature pattern. With the new values of pressure and temperature, and the specified salinity distribution, fluid densities and viscosities are calculated from the equations of state [Garven and Freeze, 1984]. These four steps are repeated until the iterations converge to a stable temperature solution.

Two-dimensional finite element grids (21 rows by 44 columns in Plate 1 and 23 by 61 in Plate 2) were developed to describe the geometry of the hydrostratigraphic units in the two cross sections (Figures 6 and 7). Formation properties such as permeability, porosity, thermal conductivity, and fluid calcced from the equations of state [Garven and Freeze, 1984].

Plate 1. (opposite) Simulation results of the OilGen code, the coupled steady state variable-density groundwater flow, and heat conduction-convection model, carried out along the A-A’ geologic cross section (Figure 6) using the estimated material properties (Table 2 and 3). Results include (a) finite element mesh; (b) hydrostratigraphy, where properties of units A-G are defined in Table 2; (c) computed equivalent freshwater hydraulic head distribution; (d) computed groundwater velocity (meters per year), where vector length is linearly proportional to velocity rate, illustrating that most groundwater discharges towards the Cabri spring in the west and Tabba springs in the east; (e) stream function exhibiting the deep, buoyancy-driven, free convection cell beneath the Golan Heights, and the shallow, gravity-driven, convection beneath the upper Galilee Mountains; and (f) temperature (degrees Celsius) distribution exhibiting the geothermal anomaly at Tabba springs because of the convective deep groundwater flow and that beneath the Golan Heights resulting from the free convection cell. Note the differences between this and the next plates, which explain the major differences in field observations.
salinity were assigned for each element (Plates 1b and 2b; Table 2). Other parameters (Table 3) are an integral part of the code. Boundary conditions for groundwater flow and heat transport were defined and are described below. Many of these parameters and boundary conditions were necessarily based on generalized assumptions and simplifications; however, the sensitivity to changes in assigned values of some of these parameters were analyses. Results included equivalent freshwater hydraulic head distribution, velocity field, stream function, and temperature distribution for each of the geologic cross sections (Plates 1c–1f and 2c–2f).

Boundary conditions were defined as follows. Levels of the groundwater table in phreatic portions of aquifers and piezometric heads at their confined portions are routinely monitored at tens of wells all over Galilee and around Lake Kinneret [Hydrological Survey, 1996]. In lower Galilee, hydraulic head reaches elevations of 20–40 m above mean sea level. In upper Galilee, it reaches levels of 180–220 m. Westward and eastward of these locations, water table gradients are different, as the base level at the west is the Mediterranean Sea, while at the east it is Lake Kinneret, where surface elevation is 210 m below sea level. Hydraulic heads in wells around the lake depend on the formation in which the well is completed or screened. At southern Golan Heights the regional water table is at about 40 m below sea level. The mean water table configurations were taken as boundary conditions at the surface of the two cross sections. For the steady state modeling, we have assumed a constant water table position (constant recharge and discharge rates through time). At the base of the cross section (4–5 km deep), at the west side (beneath the Mediterranean Sea shore), and at the east side (beneath the plateau of the southern Golan Heights), no-flow boundary conditions were assumed.

Thermal boundary conditions were defined as follows. Heat fluxes of 50 ± 24 mW/m² were measured at 70 locations in Israel [Eckstein and Simmons, 1978; Eckstein and Maurath, 1995]. Slightly elevated heat fluxes (75 ± 5 mW/m²) were measured at the lake itself (Ben-Avraham et al., 1978). These values reflect the redistributed heat fluxes induced by groundwater circulation [Gvirtzman et al., 1997]. The lower boundary conditions at 4–5 km depth reflect the undisturbed heat flux; thus a steady geothermal flux of 60 mW/m² was assumed beneath the Galilee Mountains and 72 mW/m² beneath the rift valley (no data are available to support isothermal conditions or any other heat flux distribution). Insulated boundaries have been assumed at both sides of the cross section. Constant temperature of 20°C was assumed at the surface. The assigned thermal conductivity values (Table 2) were based on those reported in the literature [Eckstein and Simmons, 1978; Eckstein and Maurath, 1995], which range between 0.8 and 3.6 W °C⁻¹ m⁻¹, depending on the specific lithology.

**Simulation Results**

Based on both sets of simulations (Plates 1 and 2), two types of basin-scale groundwater convection, and therefore of heat transfer, are suggested on both sides of the rift valley. The first is a simple forced convection which takes place at the western side of the graben. Rainwater discharges at the Galilee Mountains, flows vertically to the shallow Judea aquifer and partially to the deep Kurnub and Arad aquifers (gradually heated), and

![Plate 2](image-url) Simulation results of the OILGEN code, the coupled steady state variable-density groundwater flow, and heat conduction-convection model, carried out along the B–B’ geologic cross section (Figure 7) using the estimated material properties (Table 2 and 3). Results include (a) finite element mesh; (b) hydrostratigraphy, where properties of units A–F are defined in Table 2; (c) computed equivalent freshwater hydraulic head distribution; (d) computed groundwater velocity (meters per year) where vector length is linearly proportional to velocity rate, illustrating that most groundwater discharges towards the Na’aman spring in the west and Fulya springs in the east; (e) stream function exhibiting the deep, buoyancy-driven, free convection cell beneath the Golan Heights, and the shallow, gravity-driven, convection beneath the lower Galilee Mountains; and (f) temperature (degrees Celsius) distribution exhibiting the geothermal anomaly at Fulya springs because of the convective deep groundwater flow and that beneath the Golan Heights resulting from the free convection cell. Note the differences between this and the previous plates, which explain the major differences in field observations.

### Table 2. Hydrologic Parameters Assigned to the Hydrostratigraphic Units

<table>
<thead>
<tr>
<th>Unit* Parameter</th>
<th>Aquifer</th>
<th>Aquitard</th>
<th>Aquiclude</th>
<th>Basement, G†</th>
</tr>
</thead>
<tbody>
<tr>
<td>Formations‡</td>
<td>4c, 5b</td>
<td>1b, 2b, 3, 4b, 6b, 11</td>
<td>1a, 4a, 5a, 7</td>
<td>2a, 9</td>
</tr>
<tr>
<td>Kᵥ, m³/yr</td>
<td>200</td>
<td>10</td>
<td>5</td>
<td>0.1</td>
</tr>
<tr>
<td>Kᵥ, m³/yr</td>
<td>2</td>
<td>0.1</td>
<td>0.05</td>
<td>0.001</td>
</tr>
<tr>
<td>ρ, g/ml</td>
<td>15</td>
<td>8</td>
<td>3</td>
<td>2</td>
</tr>
<tr>
<td>λ, W/°Cm</td>
<td>3</td>
<td>3</td>
<td>3</td>
<td>4</td>
</tr>
</tbody>
</table>

*Corresponds to hydrostratigraphic units defined in Plates 1b and 2b.
†Unit G appears in cross section A–A’ only.
‡Corresponds to formations shown in Table 1 and Figures 6 and 7.
§Corresponds to formationsshowninTable1andFigures6and7.

**Table 3. Constant Parameters Used in Numerical Simulations**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Symbol</th>
<th>Unit</th>
<th>Value</th>
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</thead>
<tbody>
<tr>
<td>Reference fluid density</td>
<td>ρ₀</td>
<td>kg/m³</td>
<td>1000</td>
</tr>
<tr>
<td>Reference fluid viscosity</td>
<td>μ₀</td>
<td>kg m⁻¹ s⁻¹</td>
<td>10⁻³</td>
</tr>
<tr>
<td>Fluid thermal conductivity</td>
<td>kₚ</td>
<td>W °C⁻¹ m⁻¹</td>
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<td>Fluid heat capacity</td>
<td>cₚ</td>
<td>J °C⁻¹ kg⁻¹</td>
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</tr>
<tr>
<td>Solid heat capacity</td>
<td>cₛ</td>
<td>J °C⁻¹ kg⁻¹</td>
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<tr>
<td>Geothermal heat flux</td>
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<td>mW m⁻²</td>
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<td>°C</td>
<td>20</td>
</tr>
</tbody>
</table>

*At the rift valley, heat flux is considered 20% higher; that is, 72 mW m⁻².
discharges at the lake’s western margin, creating the hot springs. The springs are actually a mixture of fresh and salty sources emerging from these aquifers. The fresh component drains laterally from the shallow aquifer, while much of the saline component emerges as hot brines from the deeper aquifers. The model predicts that average linear velocities at the Judea, Kurnub, and Arad aquifers approach 40–100 m/yr, 5–8 m/yr, and 0.8–1.1 m/yr, respectively (Plates 1d and 2d), depending on the elevation of groundwater table at the Galilee Mountains and the hydraulic parameters of each of the hydrostratigraphic units (especially permeability and anisotropy).

A second groundwater flow mechanism exists on the eastern side of the basin. It is a free convection cell driven by buoyancy forces. A counterclockwise-flowing (on the E-W section), thermally driven convection cell, 6–10 km in size, is proposed in the confined, 3-km-thick, Cretaceous-Jurassic aquifers beneath the southern Golan Heights. Its characteristics are discussed in detail by Gvirtzman et al. [1997].

Because precise values for many of the hydrological parameters are unknown, especially at a large basin scale, the sensitivity of the modeling results was checked against variations in assigned parameters. The results shown in Plates 1 and 2 are outputs of many simulations in which parameters were modified until the comparison between the calculated and the actual recharge, discharge, and geothermal gradients seemed favorable (Table 2). The simulations estimate average recharge of 250 and 100 mm/yr at the upper (Figure 6) and lower (Figure 7) Galilee Mountains, respectively; discharges of 900 and 2300 m³/yr, and 0.8–1.1 m/yr at the upper (Figure 3 and 4). As seen in Plates 1 and 2, the waters that emerge at these springs are actually mixtures of groundwaters coming from the upper (Judea group), the middle (Kurnub group), and the lower (Arad group) aquifers. Solute concentrations measured in these aquifers differ from each other by orders of magnitude [Goldshmidt et al., 1967; Starinsky, 1974; Mercado and Mero, 1984]. The upper aquifer contains fresh groundwater with concentrations in the range of 50–500 mgCl/L. In the middle aquifer, brackish groundwater of 5000–10,000 mgCl/L is found. Concentrations in this aquifer change gradually from fresh groundwater at the phreatic portion to brackish groundwater towards the outlet [Michelson, 1975]. In the lower aquifer, brines with concentrations in the range of 50,000–100,000 mgCl/L exist [Mercado and Mero, 1984]. On the basis of the stream functions, it is possible to determine the mixing ratios of the three aquifers at the outlets and thereby to estimate the spring salinities. As a rough example, the upper fresh aquifer, the middle brackish one, and the lower saline contribute 80–95%, 5–15%, and 1–3%, respectively, of the water that emerges through the Tabba and Fulya springs (Plates 1e and 2e); thus it is consistent with the general range of concentrations, 500–3000 mgCl/L, measured at these springs. The objective of the following simulations is not to precisely reconstruct field observations, because results are sensitive to many poorly constrained hydraulic parameters. However, insight into the response of the system to changes in groundwater elevation can be gained.

Through these analyses it was found that flow rates are particularly sensitive to uncertainties of regional permeability, but the same flow patterns persist throughout the tested permeability range (10–500 m/yr for the Judea and Kurnub aquifers). In these cases, calculated (simulated) values were compared with known data (recharge, discharge, and geothermal gradients). However, in other cases, a change in an assigned permeability caused a significant change in flow pattern. For example, a reduction in hydraulic conductivity of the Jurassic aquifer to below a certain value (K_H = 1 m/yr, where K_H/K_V = 100; see Table 2) ceases fluid circulation in the free convection cell, and thus no heat anomaly is formed; this is inconsistent with field observations.

The estimated hydraulic conductivity of the Judea-Kurnub aquifer formations, K_H = 200 m/yr (Table 2), is slightly higher than previously reported. Through a detailed study, Bein [1967] reported permeabilities of 0.01–1.0 mdarcy using core permeameter measurements, and hydraulic conductivity of 1 m/d using aquifer pump tests. Using drill stem tests in these formations, Nativ and Menashe [1991] reported permeabilities of 1–20 mdarcy. However, the hydraulic conductivity of carbonate and sandstone aquifers, at the basin scale, is commonly higher by almost 2 orders of magnitude than the formation permeability measured in the laboratory on core plugs or by about 1 order of magnitude higher than permeability measured by a pumping test in the field [Garven, 1994]. Garven argued that the hydraulic conductivity measured in the laboratory reflects primary porosity; when measured at a borehole, it reflects the macroscale fracture sets; however, at the basin scale it should reflect karst systems and regional fracture networks. An anisotropy ratio of horizontal to vertical conductivity, K_H/K_V = 100, is assumed for most strata in the basin. This value is chosen to represent the inherent anisotropy within individual sedimentary beds and to account for the layered heterogeneity that exists at a finer scale than that depicted in a basin-scale representation [Freeze and Cherry, 1979].

Responses to Changes in Groundwater Table

The following simulations were conducted to gain insight into the inverse and direct relationships between discharge and salinity observed in Tabba and Fulya springs, respectively (Figures 3 and 4). As seen in Plates 1 and 2, the waters that emerge at these springs are actually mixtures of groundwaters coming from the upper (Judea group), the middle (Kurnub group), and the lower (Arad group) aquifers. Solute concentrations measured in these aquifers differ from each other by orders of magnitude [Goldshmidt et al., 1967; Starinsky, 1974; Mercado and Mero, 1984]. The upper aquifer contains fresh groundwater with concentrations in the range of 50–500 mgCl/L. In the middle aquifer, brackish groundwater of 5000–10,000 mgCl/L is found. Concentrations in this aquifer change gradually from fresh groundwater at the phreatic portion to brackish groundwater towards the outlet [Michelson, 1975]. In the lower aquifer, brines with concentrations in the range of 50,000–100,000 mgCl/L exist [Mercado and Mero, 1984]. On the basis of the stream functions, it is possible to determine the mixing ratios of the three aquifers at the outlets and thereby to estimate the spring salinities. As a rough example, the upper fresh aquifer, the middle brackish one, and the lower saline contribute 80–95%, 5–15%, and 1–3%, respectively, of the water that emerges through the Tabba and Fulya springs (Plates 1e and 2e); thus it is consistent with the general range of concentrations, 500–3000 mgCl/L, measured at these springs. The objective of the following simulations is not to precisely reconstruct field observations, because results are sensitive to many poorly constrained hydraulic parameters. However, insight into the response of the system to changes in groundwater elevation can be gained.

Since streamlines can be drawn by any given intervals (ΔΨ), it is possible to estimate the total discharge through each hydrostratigraphic unit. Starting with the geological cross section through Fulya (B-B’; Figure 7), attention was focused on a portion 25 km long by 1.5 km deep (Figure 8a). This portion contains the recharge area in the lower Galilee Mountains and the discharge area at Fulya springs, including the upper and middle aquifers only. The stream function was drawn with ΔΨ = 50 m²/yr. It can be seen that under steady state flow conditions (water table peak at 40 m), recharge into the upper and middle aquifers are 650 and 250 m²/yr, respectively (Figure 8b). While flowing eastward through these aquifers, 100 m² of water leaks from the middle aquifer into the upper one each year, owing to the hydraulic head gradient that exists across the separating aquitard (Plate 2c). Consequently, the yearly discharge at the Fulya springs is a mixture of 750 and 150 m² of water from the upper and middle aquifers, respectively. The lower aquifer supplies about 20 m² of the total spring discharge.

Additional flow simulations with a lower water table elevation (peak at 14 m) were conducted (Figure 8c) by reducing the width of the two uppermost rows of elements. Under such conditions, recharge rates into the upper and middle aquifers are 650 and only 150 m²/yr, respectively, and the seepage rate from the middle aquifer towards the upper one is only 50 m²/yr. Consequently, yearly discharge at the Fulya springs is a
mixture of 700 and 100 m² of water from the upper and middle aquifers, respectively. This numerical experiment indicated that under a declined groundwater table, discharge from the brackish groundwater aquifer is significantly smaller, while that from the freshwater aquifer has not significantly changed. Thus spring salinity is lower and discharge is smaller under conditions of lowered groundwater table. These two simulations (Figures 8b and 8c) exhibit the direct correlation between discharge and salinity observed in the Fulya springs (Figure 4).

The simulations were conducted assuming steady state conditions and ignoring seasonal changes in discharge and salinity. Nevertheless, they provide insight into the general behavior of the system and help explain field observations. These results agree with the conceptual model suggested by Goldshmidt et al. [1967], as is illustrated in Figure 2b.

The same procedure was applied at the geologic cross section passing through Tabha springs (A-A′; Figure 6). Again, attention was focused at the recharge and discharge portions at the eastern Galilee Mountains for two elevations of groundwater table. Figure 9 illustrates the steady state flow lines on a 5-km-long and 1.0-km-thick cross section, where flow converges from various aquifers towards the Tabha springs. Given a high groundwater elevation (peak at 220 m), discharges are 2200 and 100 m²/yr from the upper and middle aquifers, respectively; while under a declined groundwater table (peak at 180 m), discharges are 1700 and 100 m²/yr, respectively. Therefore, under a higher groundwater table, the freshwater source significantly dilutes the lower brackishwater source, whereas under the lower groundwater table, the brackish-water component is not significantly diluted. In other words, the peculiar behavior of the Tabha springs, the inverse correlation between salinity and discharge (Figure 3), was reproduced.

The two sets of simulations (Figures 8 and 9) produce different results because of the different hydrogeological structures of the two subsurface basins. In the Tabha drainage basin, only the upper aquifer is phreatic (unit 5b in Table 1;
Figure 6 and Plate 1b), but the lower aquifer (unit 4c) is totally confined. In the Fulya system, both aquifers (units 4c and 5b in Table 1; Figure 7 and Plate 2b) are partially phreatic. Indeed, outcrops of the Lower Cretaceous aquifer (unit 4c, namely, Nebi Said and En el Assad formations) are exposed only in lower Galilee, within the Fulya drainage basin [Golani, 1961; Eliezri, 1965]. As a result, changes in elevation of the groundwater table cause different responses in the middle aquifers. At Tabha, changes in elevation of the groundwater table are significantly expressed in the upper aquifer but are smoothed out through the aquitard separating the two aquifers. Therefore fluctuations in elevation of the groundwater table have a negligible effect on the head distribution in the middle aquifer. On the other hand, at Fulya, a change in groundwater table also affects the head distribution in the middle aquifer. To summarize, these simulations reproduce the responses to changes in groundwater table of two adjacent hydrologic spring systems expressed by the discharge-versus-salinity relationships (Figures 3 and 4).

Conclusions
The observed direct and inverse relationships between a spring’s discharge and salinity, found at the western coast of Lake Kinneret, has previously been explained using two contrasting conceptual models: compaction-driven flow at Tabha, and gravity-driven flow at Fulya. Because the two groups of springs are located only 7 km apart, drain the same aquifer systems, and contain almost the same mixture of freshwater and brine sources, the coexistence of two opposite mechanisms
is questionable. This study suggests that gravity-driven flow, that is, the leaching (or the stagnant brine) model (Figure 2b), suggested by Goldsmith et al. [1967], takes place at both systems, not only at the Fulya system, as suggested by some authors [e.g., Kahanovich and Mero, 1973; Bein, 1978; Mercado and Mero, 1984]. The different discharge-versus-salinity correlation at the two springs result from the two hydrogeologic configurations of their drainage basins: at the Tabha system the upper aquifer is phreatic and the lower one is totally confined, while at the Fulya system, both aquifers are phreatic, at least at some portions. By using detailed geologic cross sections and applying the OILGEN numerical code for solving the coupled variable-density groundwater flow and heat transfer equations, the responses of the two systems to changes in groundwater table elevations were reproduced. This research has also practical ramifications. For example, reducing the salinity in Lake Kinneret seems to be possible, at least in part, by reducing the elevation of groundwater table in the phreatic portions of the Kurnub Aquifer. In fact, future management schemes for the lake and its surrounding aquifers may be explored with additional numerical experiments. This study is based on a few generalized assumptions and simplifications because poorly constrained hydrologic data exist in this region. There is much uncertainty regarding the locations of faults, thickness of formations, properties of various lithostratigraphic units, and the three-dimensional hydrogeology. Nevertheless, this study is the first attempt to construct a comprehensive regional model of groundwater flow and heat transport across the Dead Sea Rift Valley, and we argue that a better understanding of the groundwater system behavior has been already achieved through this regional modeling. It is evident that further hydrological, geophysical, and geochemical investigations and additional transient numerical simulations are required in order to understand the behavior of the entire hydrogeological system. In any case, it is worth noting that absolute flow magnitudes cannot be proven, verified, or validated by such numerical models [Konikow and Bredehoef, 1992; Oreskes et al., 1994].

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G. Garven, Department of Earth and Planetary Sciences, Johns Hopkins University, Baltimore, MD 21218.

G. Gvirtzman, Department of Geography, Bar Ilan University, Ramat Gan, 52900 Israel.

H. Gvirtzman, Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem 91904, Israel.

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