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Transient simulations of large-scale hydrogeological processes causing temperature and salinity anomalies in the Tiberias Basin

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Abstract

Hot and salty waters occur in the surroundings of the Lake Tiberias. Transient numerical simulations of thermally-driven flow without salinity effects show that mixed convection can explain the upsurge of thermal waters through permeable faults and the high temperature gradient in the Lower Yarmouk Gorge (LYG). It turns out that by including salinity effects, the flow patterns differ from those of a purely thermal regime because heavy brines dampen upward buoyant flow and convective cells. Accordingly, the fault permeability had to be increased to restore a good fit with the measured temperatures. This further supports the hypothesis that the high temperature gradient in the LYG is likely due to fractures or faults in that area. The thermohaline simulations also suggest that the derivatives of relic seawater brines
are the major source of salinity. Deep brines leaching salt diapirs cannot reach the surface. However, the presence of local shallower salt bodies below the lake can potentially contribute to the salinity of the western spring and well waters, though in very small amount. This is in agreement with geochemical data according to which the major source of the brines of the Tiberias basin represents seawater evaporation brines. Besides being of importance for understanding the hydrogeological processes that salinize Lake Tiberias, the presented simulations provide a real-case example illustrating large-scale fluid patterns due to only one source of buoyancy (heat) and those that are additionally coupled to salinity.

**Keywords:** numerical modeling; Lake Tiberias; convection; brine; fault; heat anomaly

1. Introduction

Groundwater flow, heat and brine transport processes in large-scale systems are naturally coupled and mutually dependent. Physically, the coupling is mainly through the Darcy law, in which the buoyancy forces and dynamic viscosity depend on pressure, temperature and solute concentration (e.g. Ingebristen, 1999). The consequence of this coupling is that different system behaviors arise (Chen et al. 1990). On the basis of a linear stability analysis, Lapwood (1948) shows that, for a porous medium heated from below, free convection is triggered when the value of the Rayleigh number of the medium is higher than a critical number $Ra^{critical}$. Depending on the physical properties of the fluid and geological units, different free convective regimes develop in the form of thermal plumes or fingers (e.g. Nield, 1968). In basin systems, free convection is often related to the upsurge of hot
springs. For example, Severini et al. (1983) shows that convection with a normal geothermal gradient is capable of producing warm springs in northwestern Virginia.

When dissolved solutes are also involved, mass transport within the system is associated with the protrusion of the thermal plumes. In saline environments, this coupled flow is called thermohaline convection. Real study cases are the salt domes of the Gulf of Mexico, where upward brine flow along salt flanks occurs as the result of thermohaline convection (Evans et al., 1991). In the coastal aquifers of western Turkey, free convection in the faults induces seawater intrusion (Magri et al., 2012).

Depending on the relative importance of both sources of buoyancy (i.e. heat and salt), solute can be either stabilizing and dampen thermal convection (Diersch and Kolditz, 2002) or enhance gravity-driven flow, as in the case of sinking brine from shallow salt structures (Sarkar et al., 1995; Magri et al., 2009).

When thermohaline convection interacts with the regional flow imposed by the topography of the basin, the resulting flow is referred to as mixed convection. Mixed convection is invoked as the main process of ore formation in the McArthur Basin, Australia (Garven et al., 2001). The hydraulic conductivity of the units exerts the major control on groundwater flow, and therefore strongly impacts coupled processes. In this regard, permeable faults provide preferential pathways for mixed convection and discharge of the regional flow. Permeable faults can even allow convection within surrounding units that have a small Rayleigh number (McKibbin, 1986).

Here a numerical example illustrating the features of large-scale groundwater flow coupled to heat and brine transport in a faulted system is presented. The Tiberias Basin (Fig.1), in the Jordan Rift Valley serves as study case.
The Jordan Rift Valley is a series of rhomb-shaped pull-apart basins, one of which hosts Lake Tiberias, also known as Lake Kinneret or Sea of Galilee (Fig. 1). Brines are found in springs and boreholes at the shoreline of the Lake, as well as seepages from the lake's floor (Fig. 1). The springs can be classified in clusters according to their location and the local geological settings, as well as to their chemical properties (Table 1). The lake is a major freshwater reservoir for the whole area. Therefore, understanding the driving mechanisms endangering the lake is a crucial aspect to manage this important freshwater resource.

Previous numerical simulations based on a W-E cross-section crossing the lake (Gvirtzman et al., 1997a, 1997b) show that topography-driven flow from the Galilee and convection below the Golan coexist (i.e. mixed convection) and can explain different spring behaviors as well as the anomalous geothermal gradient of the area. Similarly, Roded et al. (2013) study the high heat flow below the Lower Yarmouk Gorge (LYG) along a N-S profile at the eastern side of the lake. However, those studies account neither for the salinity effects of relic brines in the units nor for the effects imposed by the presence of a salt dome. Furthermore, in Gvirtzman et al. (1997a, 1997b) faults are not represented numerically, whereas in Roded et al. (2013) the LYG sediments are strongly anisotropic with respect to the hydraulic conductivity to enhance upward heat migration. The studies of Rimmer et al. (1999, 2003) and Abbo et al. (2003) are more conceptual: though their transient models successfully explain the seasonal variability of the spring salinity and rate, they are limited to a maximum depth of ~900 m, and do not account for temperature effects. These assumptions implicitly exclude any type of deep large-scale flow patterns and source of solute below the Upper Cretaceous aquifers. Hurwtiz et al. (2000a) models
explain paleo temperature and the transition from a Late Pleistocene salt lake to
the actual fresh one. The most recent transient 3D models of Yechieli et al. (2011),
from the Geological Survey of Israel, investigate the effects of pumping from the
major aquifers on salinity and flow rates of the springs. Yechieli et al. (2011) refers
to Kessler's report (2011) on hydrological cells modeling and supplies different
forecast for salinity change as consequence of extensive pumping but does not
implement deep sources of the saline springs system.

The intention of this study is to provide a regional picture of the ongoing large-
scale transport processes that control the migration of relic seawater brines in the
Tiberias Basin and induce the high temperature anomalies. Adequate equations of
state (EOS) for fluid density and viscosity are implemented to account for
temperature and salinity dependencies. Furthermore, faults are explicitly modeled
using the equivalent porous media (EPM) approach to show the impact of narrow
permeable zones on the large-scale flow regimes and on the temperature gradient
of the LYG. For the first time, a salt dome based on the interpretation of actual
seismic lines is implemented in the numerical models. This setting allows
investigating whether the presence of salt structures below the lake can also
contribute to the salinity of the springs and wells.

The simulation results illustrate the calculated flow paths, as well as fluid
temperature and salinity patterns, supported by available temperature profiles of
deep wells and spring salinities. In a more general way, the presented scenarios also
illustrate the differences between large-scale fluid patterns due to only one source
of buoyancy (heat) and those that additionally combine salinity.
2. Data

The hydrological data available are mainly based on the yearly reports from the Water Authority of Israel (2012), hydrological surveys (BGR-WAJ, 2001), geophysical and numerical investigations (Yechiely et al., 2011; Kesler, 2011) and several hydrochemical studies in both Israeli and Jordanian sides of the study area, as recalled in this section. The data consist of interpreted geological structures, regional head values, salinity, temperature and flow rates of both springs and wells. They are used to define suitable boundary conditions for the numerical models as well as to qualitatively calibrate the results. Isotopes provide additional constraints on the possible flow paths and groundwater mixing.

2.1. The selected profile and its hydrogeological setting

The selected geohydrological profile starts on the western flank of the Gilboa-Arbel syncline, west to the Lake Tiberias, (Fig. 1). It crosses the western side of the lake, between onshore Fuliya and the offshore Ma'agan springs (“Fu” and “Ma”, Fig.1), which are representatives of the Fuliya brine type (see section 2.2) and the Tiberias hot springs (“Ti”, Fig.1). The profile runs through the Jordan Rift Valley, which is at that location occupied by the lake. At mid-point between the western and eastern shores, the profile turns to SE and crosses the Ha'on well (“Ha”, Fig. 1), the Lower Yarmouk Gorge (“LYG”, Fig. 1) and ends in the Cretaceous limestone outcrops of the Jordanian Ajlun Plateau. At its eastern part, the profile is nearly perpendicular to both the geological structure and equipotential lines of the Ajlun Cretaceous aquifer (El Nasser, 1991; BGR, 2001).

Along the studied profile, the groundwater levels vary from -150 m mean sea level (MSL) in the Lower Galilee to -209 m MSL close the lakeshore (Water authority
of Israel, 2012). The regional water table in the Ajlun Heights reaches elevation of
50 m MSL at the end of the profile (BGR-WAJ, 2001). In the LYG, the observed wells
are artesian (Siebert et al., 2014).

The geological assumptions used to construct the profile shown in figure 2 are
based on Inbar (2012) and current interpretation of seismic data from the late 90's
(Ben Gay et al., 1997), as recalled here. The Golan Heights and the Ajlun are
separated by the LYG (Fig. 2) and are usually described as a continuous ENE-WSW
syncline (Meiler, 2011; Roded et al., 2013). According to Shulman et al. (2004), an
ancient fault at the LYG explains the huge difference of Jurassic thickness between
Ajlun and the Golan Heights, as reflected in the representative cross-section (Fig.2).

The ~3 km thick successions of the Triassic (Benjamini et al., 2005) and ~2 km
Jurassic sequences in the Lower Galilee (Hirsch, 2005) differ significantly from the
much thinner equivalent units in the Ajlun (400 – 500 m each). The Jurassic section
east of the Jordan Rift Valley is primarily composed of limestones and marls,
whereas the Jurassic in the west also hosts volcanics and dolomite. For that reason,
only one unit represents the Jurassic at east, whereas in the west it is subdivided
into four “sub-units” (Fig. 2).

Above, the Lower Cretaceous Kurnub Group of the Galilee comprises the Tayasir
Basalt and the continental sandstone of the Hatira formation (Table 2), attaining a
total thickness of nearly 400 m (Rosenfeld and Hirsch 2005). Opposite, in the Ajlun
area, the 200 m sequence of mostly coarse-grained Kurnub sandstone indicates a
fluvial deposition system with brief marine ingresson (Abu Saad and Al Bashish,
1996) (Table 2). Accordingly, the different sediment fills significantly vary the
physical properties distribution, particularly the hydraulic conductivity (Table 2 and section 3.2).

In both regions, intercalations of limestones and marls are building most of the Upper Cretaceous sequence. In the Galilee, the Cenomanian-Turonian sequence is composed mostly of well-bedded dolomites and limestones and while the Turonian and Lower Senonian in the Ajlun is solely composed of limestone (Rosenfeld and Hirsch, 2005; Makhlouf et al., 1996). The Senonian chalks and marls and the Eocene limestones were deposited on a folded terrain, yielding thickness increases in relative position to the synclinal axis (Flexer, 1964).

Zemah-1 borehole, located a few kilometers south of the lake (“Z” in Fig. 1), reveals an outstanding thick succession of solid salt interbedded by limestones, clastics and magmatic intrusions (Marcus and Slager, 1985). This unique assemblage of units is named "Zemah Complex" (Inbar, 2012). Following the model suggested by Inbar (2012) and current reinterpretation of seismic data close to the path of the presented geological profile (Fig. 2), it seems that a salt dome is ascending along the western fault of the Jordan Rift Valley. Therefore, it is assumed that below the lake a salt body equivalent in time to the Zemah Complex is present (Fig. 2), the top and depth of which are not known.

2.2. The anomalies: salinity and temperature

Total dissolved solids (TDS) and temperatures of spring and well waters (Table 1) are strongly dependent on the seasons. TDS is highly variable, ranging from almost freshwater conditions in the LYG to highly saline outflows in the Tiberias basin. If one compares temperatures and salinity (Fig. 3), three groundwater types can be distinguished: (1) brines from deep wells, (2) springs along the coastline of
the lake and (3) fresh to brackish waters in the LYG. It can be seen that within the
LYG, groundwater is heated up at great depths, and the low TDS indicates that
mixing with deep brines plays a minor role (Siebert et al., 2014).

Ten meter below the lake bottom, time domain electromagnetic investigations
indicate that salinity distribution is not homogeneous (Hurwitz et al., 1999), varying
between 11 gL\(^{-1}\) and 22 gL\(^{-1}\). Relatively fresh groundwater is detected beneath most
of the shoreline because of freshwater advection from regional aquifers along the
margins.

Overall, very scarce data on TDS is available at depths greater than -800 m MSL.
According to the conceptual models of Abbo et al. (2003) the Cretaceous aquifers
carry saline waters with chlorinity of 15 to 18 gL\(^{-1}\) (up to 48 gL\(^{-1}\) TDS), whereas the
Eocene aquifers bear fresh water characterized by chlorinity between 0.03 and 0.3
gL\(^{-1}\) (up to 3.5 gL\(^{-1}\) TDS). This finding is in good agreement with the chlorinity-depth
profiles by Greitzer (1980). Those profiles further show salinities in deeper units up
to 120 gL\(^{-1}\) (172 gL\(^{-1}\) TDS in the Jurassic). Based on Br/Cl and Na/Cl molar ratios, the
back-calculated TDS of deep brine yield to values of more than 300 gL\(^{-1}\) (Möller et
al., 2012).

Groundwater temperatures vary strongly within the same cluster of springs
(Table 1). Particularly, in the LYG (Mukhebeh and Hammat Gadder, “Mu and “HG”
Fig. 1), the temperature difference between springs and wells that are less than 50
m apart can be more than 10 °C. Temperature vs depth profiles at Kinneret 10b
(“K10b”, Fig. 1) and Mezar (“M” in the LYG, Fig. 1) are shown in figure 4. The K10b
well displays an inversion in the temperature trend at about -850 m (MSL). In
geothermal systems, this type of anomaly is often associated with cool water
flowing through fractures in the surrounding units. Such changes of temperature, however, can also be due to the drilling process or active pumping in the wells. In general, while geothermal surveys indicate that the average conductive heat flux in Israel is rather low, around 40 - 45 mWm$^{-2}$ (e.g. Eckstein and Simmons, 1997; Shalev et al, 2013), the area of the Lake Tiberias is overall affected by heat flow higher than 60 mWm$^{-2}$, locally reaching 85 mWm$^{-2}$ (Shalev et al., 2008). Particularly, in the southeastern part of the study area, the temperature gradient inferred from deep wells in the LYG is 46 °C km$^{-1}$ (Fig. 4) which is almost twice the average geothermal gradient of the area.

Two phenomena are attributed to this anomalously high gradient (Bajjali, 1994):

1. ascent of deep heated groundwater along fractures or
2. heat flow perturbations related to magmatic intrusions. Here, the impact of a fault on the heat transport processes in the LYG is numerically investigated.

The numerical models presented here are aimed to understand whether the observed anomalies are related to faults and hydrologic regimes, rather than quantifying the impact of local anthropogenic processes.

### 2.3. Geochemical evidences for brine movement

Numerous hydrochemical and isotope investigations have been carried out in the study area in order to understand the brine movements responsible for causing the hot saline springs. Mazor and Mero (1969), Gat et al. (1969), Hurwitz et al. (2000a, 2000b), Klein-BenDavid et al. (2004) and Möller et al. (2012) discuss various formations of brines. All studies agree that the observed saline waters are mixtures of meteoric water with some ascending relic brines.
Möller et al. (2012) suggests that these relic brine(s) are derivatives of evaporated seawater due to 1000Br/Cl molar ratios of 5.7, which resembles seawater evaporated by about 40% (McCaffrey et al., 1987). The observed relic brines result from the evaporation of seawater that remained trapped in the Rift sediments and adjacent sedimentary rocks during the last Mediterranean transgression. These relic seawater brines are henceforth referred to as source brines (SB).

Different mechanisms driving the SB in the basin have been suggested, as reviewed by Simon and Mero (1992). During its ascent, the SB mix with shallow circulating freshwater. The forces that are responsible for upward flow can be induced by compaction of sediments, tectonic stresses or density variations due to geothermal sources (Mero and Mandel 1963; Mazor and Mero, 1969). The latter often implies the generation of convective cells, as shown in numerical models of Gvirtzman et al. (1997a, 1997b). The topography-driven flow imposed by the surrounding heights (e.g. Eastern Galilee, Golan, Ajlun) provides an additional force that flushes deep-seated saline water toward discharge areas (Goldshmidt et al., 1967). Here, the topography-driven flow is also called “regional flow”. The same process is referred to as forced convection in Gvirtzman et al. (1997a) or gravity-driven flow in Rimmer et al. (1999, 2003). Rimmer et al. (1999) and Abbo et al. (2003) considered variation of groundwater levels in aquifers showing that the springs discharge and salinity are sensitive to the near lakeshore boundary conditions. Kolodny et al. (1999), Moise et al (2000) and Möller et al. (2012; 2014) discuss leaching by recharge water of residual seawater evaporation brines from pockets.
While it is clear that faults play a major role in providing preferential pathways for fluid flow in the Eastern Galilee (Vengosh and Rosenthal, 1994), the presence of deep faulting in the LYG is still an open question. Beside the heat anomaly recorded there (sections 2.1 and 2.2), an additional indication pointing to active faults in the LYG is the high $^3$He/$^4$He found in the sampled waters (Tsur, 2013; Kaudse, 2014). $^3$He predominantly originates from the mantle and therefore implies that groundwater interacted with mantle fluids or volatiles rise though open fractures. Depending on the interplay between the above described transport processes and the mixing ratio with the regional flow of freshwater, different types of brines form. Two main branches of brines are distinguishable:

1. Brines along the eastern and southeastern shoreline of the lake and even south of the Yarmouk show low Na/Cl and high Br/Cl molar ratios resembling evaporated seawater. The most prominent brine is the Ha’on brine which best represents the diluted SB. A strongly diluted form of the SB appears in the springs of Hammat Gader (“HG”, Fig. 1). Hammat Gader brines are essentially weathering solutions of the basaltic cover in the eastern catchment area. These spring waters contain less than 3% of SB (Siebert et al, 2014). The waters nearby the Mezar deep well (“M”, Fig. 1) are recharged over limestones in the foot area of the Hermon Massif (north of the Golan, out of Fig.2) and are very diluted too (Siebert et al., 2014). The brines of Waqed (“W”, Fig.1) and nearby Newe Ur wells are the most southern ones that still show high Br/Cl and Mg>Ca but low Na/Cl ratios, typical of evaporated seawater.

2. Brines along the western shoreline of the lake, such as those of the Tiberias hot springs, Fuliya and Tabgha clusters (“Ti”, “Fu”, “Ta”, Fig. 1), are interpreted as
derivatives of the SB (Möller et al. 2012; 2014) because they show similarly high
Br/Cl and low Na/Cl ratios like the SB.

Fuliya and Tabgha clusters ("Fu", "Ta", Fig. 1) are affected by Tiberias types of
water ("Ti", Fig. 1), which suggests northward movement of brines along the
syncline. The Golan topography imposes an additional flow of SB below the lake,
from east to west (Stiller, 2009).

The Na/Cl of SB at the western shore of the lake is slightly enhanced, but still
much below one, indicating local leaching of evaporites below the lake. Therefore,
it is highly unlikely that the salinity of the observed springs originate from leaching
of the Zemach salt plug (Möller et al., 2012).

3. Modeling Approach

3.1. Numerical model

Transient simulations of coupled fluid flow, heat and mass transport processes
(i.e. thermohaline) are run in order to investigate the possible hydrological regimes
developing in the selected profile (Fig. 2).

The commercial finite element (FE) software package FEFLOW (Diersch, 2002)
is used to solve the partial differential equations (PDE) of thermohaline flow. The
strongly coupled system of equations is given in the appendix A. FEFLOW® fully
implements variable-density and viscosity form of the Darcy law (Eq. A.1).

Polynomial fittings ensure that fluid density (Eq. A6) is dependent on the calculated
pressure, temperature and concentration. Only the liquid phase is considered and
fluid viscosity (Eq. A7) is pressure independent.
The simulations are run over 1 Myr. A semi-implicit time discretization scheme with a maximum time step of 3 yr is used to advance the coupled equations. The simulated time interval of 1 Myr does not represent a specific geological period but allows the simulations to reach a quasi steady-state solution of the problem.

The two dimensional approach implies that all processes occur in the vertical x-z plane, thereby neglecting convection parallel to the fault plane, which is the most likely convective mode in fractured media (e.g. Simmons et al. 2008). Convection in the fault plane determines the number of springs along the surface trace of the fault (Lopez and Smith 1996), and further contributes to the migration of both heat and brine. Additional cross-sectional flow cutting the profile is driven by the regional topography that in turn controls the location of discharge areas. Furthermore, at the turning point of the profile (Fig.1 and Fig.2), it is not possible to apply any mass-balance. While these aspects cannot be accounted in a two-dimensional approach, the 2D patterns illustrated here are still useful indicators of the physical processes and hydrogeological characteristics controlling the temperature and salinity trends observed in the area. At the present state of the research, a 3D structural model of the whole study area is built which will allow investigating the additional impacts of cross-sectional flows and infer salinity mass-balance of the lake.

3.2. Structural setting and properties

The modeled cross section (Fig.2) consists of all units described in section 2.1, except for the poorly constrained stratigraphy below the Triassic at the eastern end of the profile. A basement at 6 km depth closes the model to apply heat flow boundary conditions (Fig.5).
The physical properties of each unit (e.g. hydraulic conductivity, storage, porosity and heat conductivity) are given in Table 2. The hydraulic values are mainly adapted from previous numerical investigations (e.g. Gvirtzman et al., 1997a, 1997b; Abbo et al., 2003; Roded et al., 2013), conceptual models (e.g. Rimmer et al., 1999) or pumping tests (e.g. Bergelson et al., 1998). The model differentiates the main aquifers and aquitards in the vertical direction (z). Furthermore, it also accounts for heterogeneities along the x direction of the profile, as described in section 2.1. By example, the presence of impervious basalt and clays at the base and top of the northern Cretaceous is implemented by assigning a hydraulic conductivity value lower than the one given at the southern side of the profile. The anisotropy ratio of host rocks \( \frac{K_z}{K_x} \) is 0.015, in the range of values inferred by Hurwitz et al. (2000a). By contrast, faults are isotropic, similarly to the investigations of Shalev et al. (2007). Here, the hydraulic conductivity of the faults varies between 30 myr\(^{-1} \) (Shalev et al., 2007) and 140 myr\(^{-1} \). The thermal conductivity of the units is an average estimated from the lithological descriptions given in Eckstein and Simmons (1977) and Shalev et al. (2013). The molecular diffusivity value is equal to the chloride self-diffusion coefficient used in the transient salt transport models from Hurwitz et al. (2000b).

3.3. Fault model and mesh

Different well-established approaches exist to model fractures in porous media (e.g. Blessent et al., 2014; Vujević et al., 2014). Here faults are modeled using the Equivalent Porous Media approach (EPM), i.e. permeable units extending from the basement to the top. This choice is dictated by the lack of structural inputs for fault
geometry that are required to apply other numerical approaches, such as discrete features. Faults are 40 m wide, as in the EPM models of faults in the Dead Sea basin, by Shalev et al. (2007). This fault aperture is very small compared to the kilometer-scale of the flow movements studied here. This configuration is suitable for the EPM approach as this study focuses on the migration of heat and contaminants over large spatial and geological time scale rather than the local prediction of solute exchange at the fault/unit interfaces. The numerical investigations of Abbo et al. (2003) provide an additional example of EPM approach to model mass transport in the faults of the Tiberias area.

The “Triangle” algorithm (Shewchuk, 1996) is used to build the finite element mesh. The elements have variable width and preserve the stratigraphic geometries (Fig. 5, zoom). Within the faults, the mesh resolution is approximately 10 meters, i.e. at least four nodes discretize the fault aperture in the x direction. The mesh allows to account for possible buoyant-driven flow within the faults. Element spacing grows gradually from the fault flanks to 50 meters in the surrounding units and basement. This spatial smoothing ensures that (Yang, 2006) : 1) the fault width is in the same order of magnitude of the elements size at the interface, which satisfies the EPM requirements and 2) diffusive-dispersive processes are correctly simulated at the matrix/fault interface.

The whole profile comprises approximately 150,000 triangular elements satisfying the Delaunay criterion. Finer meshes did not affect the calculated patterns.
3.4. **Boundary conditions (BC) and initial conditions (IC)**

The set of equations describing the thermohaline problem (Appendix A) is solved with respect to the primary variables hydraulic head \( h \), temperature \( T \) and concentration \( C \). The boundary conditions are illustrated in figure 5.

- \( h \): Based on the data provided in section 2.2, the head is set as follows:
  - At the top: a constant head (i.e. Dirichlet) is set along the eastern and western sides of the lake. In the Lower Galilee, the regional water vary between -150 m mean sea level (MSL) at the northern ending of the profile to -209 m MSL close the lakeshore. The water table in southern ending of the profile is equal to 50 m MSL.
  - In the LYG, \( h \) is set slightly above the local topography in order to simulate artesian conditions of the area.

  Along the lakebed, a transfer boundary condition (i.e. Cauchy, Eq. (1)) accounts for lake and groundwater interactions in the form a Darcy flux \( q(t) \) given by:

\[
q(t) = L(h_{ref} - h)n \tag{1}
\]

where the reference head \( h_{ref} \) is -210 m MSL (average lake level) and the coefficient of leakage \( L \) is \( 4 \times 10^{-2} \) yr\(^{-1} \) as estimated by Stiller et al., (1975). \( n \) is the vector normal to the top surface of the profile.

No groundwater flow (i.e. Neumann) is allowed through the lateral boundaries of the profile.

- \( T \): At the top, a heat transfer (i.e. Cauchy) with a reference temperature \( T_{ref} \) of 20°C is assigned. This open boundary condition (Eq. (2)) allows heat outflow through the surface, governed by

\[
q_T(t) = \theta_T(T_{ref} - T)n \tag{2}
\]
where the heat transfer coefficient $\varphi_T = 0.13 \text{ Wm}^{-1}\text{K}^{-1}$ corresponds to the heat conductivity of the Quaternary sediments (Table 2) divided by the representative element height (10 m). Accordingly, the calculated temperature $T$ can increase at the locations where thermal springs are predicted by the model.

At the basement, previous simulations used variable heat fluxes ranging from 60 mWm$^{-2}$ to 72 mWm$^{-2}$ (Gvirtzman et al., 1997a) or 50 mWm$^{-2}$ to 100 mWm$^{-2}$ (Roded et al., 2013) over different portions of the studied profiles, in order to reflect the observed thermal anomalies of the area. Since the numerical investigations presented here mainly focus on the impact of faults on redistribution of heat and brine in the system, a constant heat flux is set along the whole profile basement. Specifically, an undisturbed basal geothermal flux (i.e. Cauchy) of 60 mWm$^{-2}$ is set at the bottom of the model. This allows inferring to which extent the observed temperature and salinity anomalies result from hydrogeological features of the basin, such as circulation of thermal waters through faults, rather than imposed deep crust anomalies.

Lateral boundaries of the cross-section are insulated (i.e. Neumann).

- $C$: A Cauchy mass transfer boundary condition (Eq. (3)) is set at the top, analogously to the head and temperature boundary conditiona (Eq. (1) and Eq. (2)). The mass flux $q_C(t)$ is given by

$$q_C(t) = \varphi_C (C_{\text{ref}} - C)n \quad \text{Eq. (3)}$$

The concentration reference $C_{\text{ref}}$ is 0.22 gL$^{-1}$ at the lake (average lake salinity) and 0 gL$^{-1}$ along the remaining portions of the top profile. The mass coefficient transfer $\varphi_C$ is $4e^{-2}\text{ yr}^{-1}$. 


A constant concentration of 300 gL⁻¹ is set at salt dome (UZC). FEFLOW® cannot account for dissolution of salt and therefore the shape of the UZC does not change with time.

The transient simulations are initiated as follows (Fig. 5):

- **h**: The initial conditions for hydraulic head are derived from steady state simulations of groundwater flow.
- **T**: As for the hydraulic head, the initial temperature profile is purely conductive (Fig. 5, top). Because of the high thermal conductivity of the UZC (Table 2), isotherms are slightly bending at the salt edges. This phenomenon is very common in geothermal basins hosting salt domes (e.g. O’Brien and Lerche, 1988).
- **C**: In the paragraph 4.2, thermally-driven flow is coupled to brine transport. For this purpose, based on the data and considerations described in sections 2.2 and 2.3, an initial salinity condition is set to reflect the presence of the relic source brines (SB) resulting from seawater evaporation, as illustrate in figure 5 (bottom): the concentration increases from freshwater conditions within the top aquifers (Eocene) to 250-300 gL⁻¹ in the basement and UZC. This assumption certainly does not reflect paleo-salinity conditions of the area nor correctly represents the Plio-Miocene SB. Nevertheless, it allows studying the impact of faults, heat and regional flow on the migration of the relic seawater.
4. Results and discussions

Models of fluid transport processes over basin-scale and geological time periods like those presented here cannot be calibrated and verified for temperature and salinity at a given time and location. Nevertheless, the results are qualitatively fitted to the temperature-depth profiles of the wells and the salinity ranges of the springs described in section 2, by testing different fault permeability. This kind of “regional calibration” allows investigating the major hydrogeological processes that control the migration of heat and brine in the system.

4.1. Scenario 1: Effects of geothermal flux on flow and temperature anomalies

Following previous numerical studies of the area (e.g. Gvirtzman et al., 1997; Roded et al., 2013), only equations A.1 to A.3 are solved here, i.e., no brine transport is computed. Accordingly, fluid density and viscosity (Eq. A.6 and Eq. A.7) are not dependent on the concentration C. This simplifying assumption is made to directly compare the findings with those from previous studies and later infer the impact of salinity (section 4.2).

Velocity and temperature fields resulting from the coupled fluid flow and heat transport process are illustrated in figure 6 (top). In this scenario, all faults except the minor one below the lake are permeable, with a hydraulic conductivity of 30 myr$^{-1}$, like in the simulations from Shalev et al. (2007). Two major independent flow fields can be distinguished:

1. A topography-driven flow (or regional flow) below the Lower Galilee and southeastern heights discharges groundwater through the Turonian/Cenomanian and Upper Eocene units, respectively. Therein, the velocity ranges between 0.4 myr$^{-1}$ to 1 myr$^{-1}$. Springs exhibit peak velocities between 3 myr$^{-1}$ and 5 myr$^{-1}$ at the fault
traces that border the lake and in the discharge area of the LYG. A sensitivity analysis showed that springs flow rate per meter width reaches 3600 m³ yr⁻¹ when the hydraulic conductivity of the elements at the top of the fault is 400 myr⁻¹ (e.g. gravel-sand). Assuming springs to discharge over a km-long shoreline, the inferred spring fluxes are in the same order of magnitude of the monitored ones.

(2) Different deep-seated convective flows are separated from the upper regional flow by the major aquitards. A squeezed cell develops in the Jurassic Nirim ("JN", Fig. 6) within the northern part of the profile. The calculated Rayleigh number of the Jurassic Nirim is much smaller than the critical value of the onset of thermal convection, calculated according to the theory of Nield (1968). However, in the presence of permeable faults, it is known that convection occurs also in units with subcritical Rayleigh properties (McKibbin, 1986). Channeled buoyant flow in the faults induces groundwater in the surrounding units to flow either in a convective-like mode or directly toward the faults, like in the overlying Jurassic Zohar ("JZ", Fig. 6). Darcy velocities of a centimeter per year characterize this cellular regime, which is one to two orders of magnitude weaker than the upper regional flow. In contrast to the fault-induced groundwater flow in the Jurassic units at the northern side, thermal buoyant forces generate Rayleigh convective patterns in the LYG and Golan Heights. Two cells circulate groundwater below the Senonian aquitard at a maximum velocity of 0.1 myr⁻¹. This is due to the presence of more permeable and thick cretaceous horizons at depths which Rayleigh number is higher than the estimated Ra_critical. As a result, the heat flow destabilizes the fluid density: the vertically elongated cell below the Golan forces groundwater to descend from the Upper Turnonian units (-1 km depth MSL) to the Lower Jurassic (-3 km depth MSL),
while deep groundwater rises mostly through the LYG fault and partly through the Ha’on fault, at velocities close to 2 myr\(^{-1}\). In this respect, deep-reaching faults provide the only hydraulic connection between the deep convective systems and the shallower regional flow, allowing thermal water to ascend from depths of -3 km MSL. Discharging springs result from the interaction of these two regimes supporting the chemical data that spring waters are a mixture of deep thermal water and shallow groundwater.

The regimes described above strongly affect the temperature distribution (Fig. 6, bottom). At the northern part of the profile, the thermal water ascending to the surface along the fault flanking the lake generates an elongated heat plume. Hot groundwater is drained from the Jurassic units and flows out of the system at temperatures between 50 °C and 60 °C. The tip of the heat plume spreads also laterally toward the northern side of the profile because of the presence of open faults that partly capture the ascending flow (Fig. 6, zoom). Therefore, groundwater temperature can either vary of several degrees over the lateral temperature gradient or decrease with depth. A simulation in which buoyant forces are not computed (i.e. \(\rho_f \neq \rho_0\) in Eq.(A.2)) prevents any heat plume to ascend from depths, but reveals a local regional flushing through faults (i.e. advection) of groundwater from Upper Turonian/Eocene formations, allowing a maximum spring temperature of 32 °C. Both calculated temperatures are in agreement with the monitored spring temperatures (Table 1), suggesting that thermal buoyant flow in the faults and advection by regional flow are both possible heat transport mechanisms.

Below the LYG, the two convective cells generate an upwelling of deep groundwater into the LYG fault, and a downwelling of colder water in the fault
below the Golan. Spring temperature in the discharge area of the LYG is 35 °C owing to the presence of a strong regional flow in the upper Eocene unit that cools down the rising thermal plume. The small upwelling observed at Ha'on is due to the local groundwater outflow from the neighboring Golan, as also indicated by a spring temperature of 30 °C.

The temperature-depth profiles along two boreholes, as located in Fig.6, show a good fit with available well data (Fig. 7, red crosses and circles). The temperature inversion observed at K10b well (red circles) is due to the spreading of the heat plume toward neighboring permeable faults, as previously explained (Fig. 6, zoom). If the fault at the western side of the K10b well is impervious, the heat plume cannot spread laterally, leading to a linear vertical temperature gradient (Fig. 7, yellow circles).

In the LYG (Fig. 7, red crosses), the presence of the fault allows the temperature gradient to be steep even under normal basal heat flow conditions. Interestingly, without the LYG fault, the convective plume persists and the temperature trend is preserved (Fig. 7, yellow crosses). However, since hot water is not anymore channeled upward, the calculated temperatures are lower compared to the case with fault (Fig. 7, red crosses).

Gvirtzman et al. (1997) models display similar patterns below the Golan. The simulations of Roded et al. (2013) display a thermal plume discharging upward in the LYG. However, those results are inferred from simulations that do not account for salinity effects, as discussed in the next section.
4.2 Scenario 2: coupling with salinity

Here the fully coupled system of flow, heat and mass (i.e. salt) transport processes is solved (Eq. A.1 to A.7). The EOS account for pressure, temperature and salinity effects. It is worth recalling that an initial salinity distribution is set to model the SB that originally saturates the units. It is assumed that the TDS of the SB increases from freshwater conditions at ground level, to 300 gL\(^{-1}\) at the salt dome (Fig. 5 bottom, and section 3.4).

When the hydraulic conductivity of the faults is 30 myr\(^{-1}\) (as in the purely thermal simulations, paragraph 4.1), this initial salinity distribution overwhelms the convective regimes in the deep units and prevents any thermal buoyant flow in the faults (i.e. no thermal plume) that was previously observed in the scenario without salinity (section 4.1). Salinity also increases the dynamic viscosity of the brine (Eq. A.7) and therefore reduces the effective hydraulic conductivity of the sediments (Eq. A.5). As a result, the only observable process is the topography-induced flow that flushes SB at the lake shores and through the LYG.

A sensitivity analysis reveals that by increasing the hydraulic conductivity of the faults to values ranging between 90 myr\(^{-1}\) to 140 myr\(^{-1}\) triggers buoyant flow and best fits the measured temperature data (Fig. 7, green crosses and circles). The results of this scenario are illustrated in figure 8 and can be compared with figure 6 (no salinity scenario, section 4.1). The western fault flanking of the Jordan Rift Valley and the LYG fault remain the preferential pathways for upward flow of thermal water. While the flow patterns of the topography-driven flow in the upper units is qualitatively similar to the previous case (section 4.1- point 1), the deep convective regimes (section 4.1- point 2) are different. The cell below the Galilee Mountains
develops into the Upper Jurassic Zohar ("JZ", Fig. 8), creating a wide stagnant zone below it (Fig. 8). Hence, thermal waters that ascend along the western fault of the Jordan Rift Valley originate at depths between -2 km MSL. Also at the southern side of the profile, the cellular patterns are different from those calculated in a purely thermal regime (compare Fig. 8, top and Fig. 6, top). The flow below the Ajlun is not anymore convective but directed toward the LYG fault. Groundwater in the overlying Turonian unit is not fully drained by the Ha'on fault and can flow westward below the lake into the discharging fault. Darcy velocities characterizing this movement into the sedimentary fill are however very slow, at maximum 1 cm yr\(^{-1}\). Geochemical evidences, as explained in section 2.3 (point 2), also request this brine movement. In its lower part, the cell stretches toward the salt flank where downward flow is dominant.

Salinity and temperature distributions resulting from these hydrologic regimes are illustrated in figure 8 (bottom). The evolution of the SB can be inferred by comparing Fig. 5 (bottom) and Fig. 8 (bottom). The steep syncline structure of the units in the Lower Galilee enhances gravity-driven flow, which flushes the relic brines. Overall, wide areas of diluted SB bearing less than 0.1 g L\(^{-1}\) TDS characterize the Upper Cretaceous units in direct relation to the topography-driven flow below the major Heights. Salinities are 2.5, 6 and 1.5 g L\(^{-1}\) at the western shore spring, Ha’On well and LYG springs, respectively. The concentration profile exhibits areas of diluted brine at the eastern shoreline, because of the inflow of freshwater from the surrounding regional flow, as observed by Hurwitz et al. (1999). Accordingly, by decreasing the hydraulic conductivity of the cover basalt, the salinity near Ha’on increases to the monitored values (Table 1.).
Brine migration is strongly coupled with the heat transport in the faults. As explained in the previous section, neighboring open faults can capture ascending thermal waters (Fig. 6, zoom). As a result, the brine plume at the northwest side of the lake spreads 2 km inland over the lateral temperature gradient (Fig. 8, zoom). Therein salty groundwater is a mixture of relic seawater (SB) flushed by the freshwater regional flow through the northern Turonian units and buoyant thermal brines that reach the surface through the faults. Those brines likely contain additional fingerprints of dissolved evaporites lifted up either through the fault adjacent to the diapir crest or transported by the S-N flow above it (Fig. 8, top). By contrast, on the opposite side of the diapir flank, gravity-driven flow constrains heavy brine into the deep Jurassic/Triassic units. The thermal plume in the LYG has the potential to drive relic SB from Turonian/Senonian units toward the gorge. This upward flow is further enhanced by the regional flow from the Ajlun. As a result, the plume is narrow (i.e. diluted) and the isotherms at the southern side of the profile are flat compared to those resulting from the purely thermal regime (Fig. 6).

Contrarily to the previous scenario, without the LYG fault no thermal plume develops: heavy brines endure a lateral/downward migration at depths and the isotherms are close to the conductive regime.

**Impact of salt diapir and faulted lakebed**

A simulation initiated with freshwater conditions (i.e. 0 gL⁻¹ everywhere except the salt dome) is run in order to infer the impact of brine diffusion from the salt diapir. The results show that no brine reaches the surface. Only a light salty plume above the salt crest diffuses below the western shore of the lake with maximum TDS of 0.2 gL⁻¹, which supports the hydrochemical evidence of leached evaporites in the
wells along the western side of the lake (Möller et al., 2012) and not in Ha’On well.

The presence of a permeable fault crossing the lakebed does not strongly affect the
large-scale patterns of heat and salt transport within the profile. Only a local effect
on the salinity distribution below the lake is observed. The calculated concentration
of the offshore spring is 6.7 gL⁻¹, and its temperature is 22 °C. Because of the
westward flow in the sedimentary fill, additional SB and salty water diffusing form
the salt crest can reach onshore springs. Consequently, the calculated concentration
of the western spring increases to 3 gL⁻¹.

The 2D limitation of these models prevents the assessment of additional brine
and heat flow from cross-cutting faults, such as from Tiberias to Fuliya, and
transversal regional flow. Also, the models do not account for overpressured
aquifers that can further increase upward heat and brine migration. These two
aspects could explain the extremely high salinity of the Tiberias springs (Table 1),
which cannot be reproduced by the simulations.

5. Summary and conclusions

Available hydrochemical data indicate that the thermal springs in the Tiberias
Basin (TB) mainly discharge deep waters that represent diluted source brines (SB),
resulting from evaporation of seawater of the last Mediterranean transgressions.
Geothermal data and geological considerations on the TB suggest that fault-
controlled hydrothermal processes generate the anomalous temperature
inversions observed at depths and the upsurge of hot/saline spring waters. The
variable salinity of the springs is due to the changing rates of the freshwater
topography-driven flow (regional flow).
First, numerical simulations of coupled heat and flow processes (without salinity effects) are run over a representative geological profile of the area to gain insights into heat transport in the system (section 4.1, Fig.6). The hydrothermal behavior of the system is the expression of two different large-scale flow patterns separated by the Cretaceous aquitards (Table 1). (i) A topography-driven flow rapidly discharges groundwater at lowlands. In this regional movement, groundwater gains heat while flowing through the Turonian and Eocene units. The springs temperature is around 30°C. (ii) The underlying convective regimes are either induced by ascending hot waters in the faults or by density instabilities in the thick permeable Cretaceous units below the LYG. Both types of cells are observable under an undisturbed basal heat flow regime of 60 mWm², which is a representative value for the study area. Buoyant flow of thermal waters in the faults can explain temperature higher than 30°C observed at some springs. In this regard, the results are in good agreement with available data (Fig. 7, red). Temperature inconsistencies recorded at the same depth of different wells may reflect the diverse hydrogeological behavior of the surrounding faults. Besides permitting upsurge of hot water, permeable faults induce radial temperature gradients by partly capturing upward groundwater flow (e.g. K10b well), or by allowing recirculation of cool groundwater from shallower units, as in the fault below the Golan Heights (Fig. 6, zoom). It turns out that in a freshwater environment the thermal plumes in the LYG form even without a fault. This is due to the convective downwelling of cool water in the Golan Heights that pushes hot water toward the LYG.

In the thermohaline simulations (section 4.2), an initial salinity representing evaporated seawater brine, the source brine (SB), is implemented (Fig. 5, bottom).
It turns out that only the regional flushing in the shallow units is observable and the deep heat transport is close to conduction, overwhelmed by the initial salinity distribution. Topography-driven flow alone cannot support the anomalous temperature profiles observed in the wells nor springs temperature above 30°C. The anomalies are numerically reproduced by further increasing the permeability of the faults (Fig. 7, green and Fig. 8, zoom). Besides transporting heat, permeable faults permit the regional flow to dilute and flush the SB. In this regard, the hydraulic conductivity of the Quaternary sediments and the basalt cover also plays an important role on the discharge rates of the springs. The springs of Hammat Gadder in the LYG discharge highly diluted SB. By contrast, at depth below -2.5 km MSL, areas of highly saline and quasi-stagnant SB, as well as downward gravitational flow from the salt body, characterize the whole profile.

Northwestward groundwater flow may have the potential to transport small amount of salt from shallow salt bodies below the lake. Instead, no brine plume from salt diapir reaches Ha’on, at the eastern side of the lake.

To some extent, the presented results are illustrative of the present day situation because of the lack of paleo data and the static nature of the structural features of the model. Nevertheless, the results provide the basis to extend the models to a three dimensional scenario. 3D models of selected areas of the TB are currently being build and will allow studying the different convective modes in the fault planes, that can explain the complex brine movements along the master faults.

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References


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The Tiberias Basin (TB). Study area including: location of the modeled cross section (Fig. 2), topography and lake bathymetry (SRTM data, Reuters et al. 2007), major faults (Ben-Avraham et al., 1996; Hurwitz et al., 2002c; Reznikov et al., 2004), clusters of springs, wells, deep boreholes, equipotential lines mean sea level (Water Authority of Israel, 2012; BGR 2001) and suggested groundwater flow directions (Bergerlson et al. 1998). LYG: Lower Yamourk Gorge.
Location | Abbreviation | TDS (g L⁻¹) range | Temperature (°C) range
--- | --- | --- | ---
Tabgha (Cs) | Ta | 2.25 – 5.23 ¹ | 19 – 29 ¹
Fuliya (Cs) | Fu | 2.06 – 2.72 ¹,⁷ | 27 – 30 ⁴,⁷
Tiberias Hot Spring | Ti | 28.94 ¹ | 64 ⁴
Mukhebeh (Cs) | Mu | 0.5 ⁹ | 33 – 43 ⁹
Hammat Gader (Cs) | HG | 0.64 – 1.22 ¹,⁹ | 28 – 50 ¹,⁷
Gofra (Cs) | Go | 5.07 ⁷ | 32 ⁷
Hiltin 3 (w) | H3 | 0.48 – 0.517 ⁶,⁷ | 25.8 ⁷
Kinnerer 10 (b) | K10b | 24.7 – 31.7 ⁷ | 46 – 52 ⁷
Ha’on (w) | Ha | 14 – 22.5 ¹,⁷ | 24 – 35 ¹,⁷
Zemah-1 (b) | Z | 220 – |

**Table 1** Range of temperatures, Total Dissolved Solids (TDS) and flow rates for the major cluster of springs (Cs), wells (w) and boreholes (b) as located in figure 1. The values are adapted from selected publications (superscript number) and do not provide a strict minimum-maximum interval.


**Fig. 2.** Representative NW-SE cross section from Upper Galilee (Israel) to the Ajlun (Jordan) including wells and Lower Yarmouk Gorge (LYG) locations. Vertical exaggeration 3:1. Based on Saltzman et al. (1964), BGR (1993), Inbar (2012), Meiler (2011). In the southern ending of the profile, question mark symbols (?) refer to geological features that remain poorly constrained and are not implemented in the numerical model. The presence of a fault in the LYG is still debated. The physical properties of the different units and their values are listed in Table 2.
<table>
<thead>
<tr>
<th>Epoch</th>
<th>Unit abbreviation</th>
<th>Segment of the profile (NW - SE)</th>
<th>Formation</th>
<th>Lithology</th>
<th>kx,y (m/yr)</th>
<th>storage 10^6 (1/m)</th>
<th>Poro (I)</th>
<th>Heat cond (W/m/K)</th>
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<td>3.0</td>
<td>0.15</td>
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<tr>
<td></td>
<td>CE</td>
<td>SE</td>
<td>Dier Hanna</td>
<td>limestone, dolomite interbedded with marls - aquitard dolomite (some marls) - aquifer marly limestone - Aquifer dolomitic limestone - Aquifer</td>
<td>50</td>
<td>2.0</td>
<td>0.1</td>
<td>2.3</td>
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<td>Lower</td>
<td>LwC</td>
<td>NW</td>
<td>Kurnub Gr., Hatira, Taysar basalt</td>
<td>Basalt and sandstone mostly permeable sandstone</td>
<td>6</td>
<td>1.6</td>
<td>0.08</td>
<td>2.2</td>
</tr>
<tr>
<td>Jurassic</td>
<td>Zohar, Haluza, Sderot</td>
<td>NW</td>
<td>Basalt and sandstone mostly permeable sandstone</td>
<td>limestone and dolomite</td>
<td>10</td>
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<td>0.05</td>
<td>2.8</td>
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<td>Rosh Pina</td>
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<td>limestone and dolomite basalt and pyroclastics</td>
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<td>0.6</td>
<td>0.03</td>
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<td>2.2</td>
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Table 2 Units, stratigraphy and values of the physical parameters of the modeled units. The stratigraphy is based on Inbar (2012) and BGR – WAJ reports (1993). The assigned values are adapted from Gvirtzman et al. (1997a, 1997b), Bergelson et al. (1998), Shalev et al. (2007, 2013), Roded et al. (2013).

Fig. 3 TDS vs surface temperature correlations for deep wells (1), springs along the lakeshore (2), LYG springs and wells (data after Möller et al., 2014). For locations, refer to figure 1. Qiqar well belongs to the Tiberias cluster of springs.
Fig. 4. Temperature profile in the Mezar (M) in the LYG (Roded et al. 2013) and in the Kinnereth 10b (K10b) boreholes (Michelson et al., 1995). The studied profile crosses the LYG at approximately -150 m msl, K10b wellhead is at -208 m m.s.l. Linear interpolations (dashed lines) do not represent the actual temperature trends. The geothermal gradient for the Kinnereth borehole is estimated by Gvirtzman et al. (1997b).
**Fig. 5**: Boundary and initial conditions. **Top**: Initial temperature (T) distribution (filled contours, in °C) with applied flow and temperature boundary conditions (blue and red lines). Vertical exaggeration 3:1. The local zoom (no vertical exaggeration) shows the triangular finite-element mesh in the fault (40 m wide) and surrounding units. Local refinements ensure that at least four elements discretize the whole fault in the X direction. **Bottom**: Initial TDS (C) distribution (filled contours, in g L⁻¹) with applied mass boundary conditions (green lines). This salinity profile initiates the transient thermohaline simulations described in paragraph 4.2. It represents an imaginary source brine (SB) resulting from relic evaporated seawater that saturates the units. Salinity constraints are based on the geochemical data given in paragraph 2.2. Vertical exaggeration 3:1. For units abbreviations, refer to figure 2 and Table 2.
Fig. 6: Scenario 4.1 without salinity effects. Top – Colored patterns represent the velocity field (Darcy flow in meter per year) and green dashed lines indicate flow paths (no streamline). Arrows show flow direction. No flow lines are plotted in low velocity field (i.e. lower than 1 cm yr⁻¹). Bottom – Calculated temperature profile (°C). Green vertical lines locate two boreholes based on the depths of the Kinnereth 10b (K10b) and Mezar (M, in the LYG) deep wells. The temperature-depth profiles of these two boreholes are illustrated in Fig. 7. The zoom shows the lateral spread of the heat plume due to the presence of a parallel permeable fault.
Fig. 7. Calculated temperature-depths at northern (K10b, circles) and southern (M, crosses) sides of the profile compared to field data (squares, Fig. 4). The boreholes are located in Fig.6, bottom and Fig. 8. These putative boreholes are just an illustrative example based on real depths, to show the different temperature trends. Red: scenario in which mass transport is not computed (i.e. no salinity effects, section 4.1). The hydraulic conductivity of the faults is 30 m yr⁻¹. Green: scenario with brine transport (section 4.2) is 140 m yr⁻¹. Yellow: impact of a closed fault near K10b and absence of the fault in the LYG, no buoyancy from salinity effects.
Fig. 8: Scenario 4.2 showing the effects of salinity. **Top** – Colored patterns represent the velocity field (Darcy flow in meter per year) and green dashed lines indicate flow paths (no streamline). Arrows show the flow direction. No flow lines are plotted in low velocity field (i.e. lower than 1 cm yr⁻¹). Small arrows in the sedimentary fill below the lake further highlight the very low velocity field. **Bottom** – Calculated TDS (g L⁻¹) of the source brine distribution and temperature profile (°C, red dashed lines). Heavy brine from the salt diapir (UZC) remains deep-seated, as explained in the section “Impact of salt diapir and faulted lakebed”. Pink vertical lines locate two boreholes based on the depths of the Kinnereth 10b (K10b) and Mezar (M, in the LYG) deep wells. The temperature-depth profiles of the two boreholes are illustrated in Fig. 7. The zoom shows the lateral spread of the heat and brine plume due to the presence of a parallel permeable fault.