2D and 3D coexisting modes of thermal convection in fractured hydrothermal systems - implications for transboundary flow in the Lower Yarmouk Gorge

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Abstract

Numerical investigations of 2D and 3D modes of large-scale convection in faulted aquifers are presented with the aim to infer possible transport mechanisms supporting the formation of thermal springs through fractured aquicludes. The transient finite elements models are based on idealized structural features that can characterize many hydrothermal systems. The sensitivity analysis of the fault permeability showed that faults cross-cutting the main regional flow direction allow groundwater to be driven laterally by convective forces within the fault planes. Therein thermal waters can either discharge along the fault traces or exit the fault through adjacent permeable aquifers. In the latter case, the resulting flow is helicoidally and transient. The location and the spacing between discharge areas can migrate with time, is not strictly constrained to the damage zones and reflects the wavelength of the multicellular regime in the fault zone.

An illustrative example based on a simplified structural data of the Lower Yarmouk Gorge (LYG) is presented. The numerical calculations indicate that crossing flow paths result from the coexistence of fault convection, developing for example along NE-SW oriented faults within the Gorge, and additional flow fields. The latter are induced either by topography N-S gradients, e.g. perpendicular to the major axe of the Gorge, or by local thermal convection in permeable aquifers below the Eocene aquiclude.
Sensitivity analysis of fault hydraulic conductivity ($K$) and the analytical solutions based on viscous-dependent Rayleigh theory show that $K$ values between $2.3e^{-7}$ m/s and $9.3e^{-7}$ m/s (i.e. 7 m/yr and 30 m/yr, respectively) favor coexisting transport processes. The uprising thermal plumes spread over several hundred meters forming clusters of springs, in agreement with observation, and which temperature fall within the measured ranges, i.e. 20°C –60°C. To some extent the models also reproduced the transient behavior of the spring temperature. Owing to the idealized nature of the presented models, the numerical results and the associated analytical solution can be applied to study the onset of thermal convection and resulting flow patterns of any fractured hydrothermal basin.

1. Introduction

In most hydrothermal systems, the dependence of fluid density on temperature leads to unstable density stratifications in which a colder (denser) fluid overlies a warmer (lighter) fluid (Nield and Bejan, 2006). The resulting convective flow is also referred to as buoyancy-driven flow or thermally-induced flow. Since density differences can extend over large spatial scales and persist over geological time periods, thermal convection is often invoked to explain natural processes in sedimentary basins such as dolomitization (Gasparrini et al., 2006), hydrothermal mineralization (Kühn M. & Gessner K. 2009; Harcouët-Menou et al., 2009), seawater intrusion (Magri et al., 2012) or CO$_2$ dissolution (Bickle, 2009).

Early studies of the problem highlight that the onset of thermal convection in the Earth crust requires a relatively high permeability of the rocks (Lapwood, 1948). Accordingly, thermally-induced flow develops in permeable faults or fractures. Analytical calculations from Malkovsky and Pek (2004) show that, under a normal geothermal gradient (0.025 °C/m), a damage zone of 100 m width embedded in an impervious rock allows for convection already at hydraulic conductivity of $1e^{-7}$ m/s. Several numerical models suggest that thermal convection in faulted basins can explain the spatial correlation between fractured zones and heat flow anomalies or the occurrence of hot springs (e.g., Ormond et al., 1995; Rabinowicz et al., 1999; Person et al., 2012; Schilling et al., 2013; Kaiser et al., 2013). These studies further reveal that
thermal convection in a faulted system generate complex flow patterns either i) within the faults (finger-like patterns) or ii) across the neighboring permeable aquifers (convective roll or helicoidal patterns).

An example in which these modes of convection might coexist is the Lower Yarmouk Gorge (LYG), at the border between Israel, Jordan and Syria (Fig. 1). In the LYG, thermal waters ascend through fractured aquicludes forming hot springs. The latter occur over wide areas forming clusters that are close to or along the major fault axis. Groundwater temperatures within the same cluster of springs (Fig. 1) strongly vary. Particularly, in the LYG clusters, the temperature difference between springs or artesian wells that are less than 50 m apart can be more than 10 °C.

![Figure 1. Clusters of thermal springs (red triangles) occur all around Lake Tiberias and within Lower Yarmouk Gorge (LYG). Multi-directional groundwater flow from heights and buoyancy forces through faults generate the Lower Yarmouk Gorge (LYG) cluster of springs. Each cluster represents wide areas where different thermal springs are observable.](image)

Hydrochemical studies from Siebert et al. (2014) indicate an intense anisotropic hydraulic functionality of the lower gorge that allows for both channeled groundwater flow along the longitudinal axle (E-W) and cross-cutting nearly meridional flow.

Previous 2D models from Magri et al. (2015) suggested that the LYG hot springs (Fig. 1) are a mixture of groundwater from the surrounding highlands (topography-driven flow) and thermal fluids ascending along faults (buoyancy-driven flow). However, the two-dimensional limitations of the simulations did not allow to fully capture convective patterns that cross-cut the major flow direction nor those that may
develop within the fault plane. Here an idealized 3D model of thermal convection in a faulted system is
built with the aim to study the development of different modes of convection, likely responsible of
upsurge of thermal water. Owing to the generalized structural features of the studied domain, the
results presented here illustrate both 2D and 3D modes of thermal convection that can characterize any
faulted hydrothermal systems and are not strictly related to the LYG. The interactions between different
convective modes in 3D unveil new mechanisms supporting the existence of complex transboundary
flow paths across faults, to date only traceable with hydrochemical investigations (e.g. Siebert et al.,
2014). This article further applies recently developed formulas to predict the onset of convection of
temperature-dependent viscous fluids in 3D faults (Malkosvky and Magri, 2016).

First, the numerical implementation applied to both 2D and 3D models is described. Then 2D and 3D
flow and heat patterns are compared. The concluding remarks about the inferred flow paths outline
general implications for thermal water migration in faulted hydrothermal systems, such as the LYG.

2. Numerical implementation

The equations governing coupled fluid flow and heat transport processes are solved using the finite
element software FEFLOW®. The mathematical formulation is given in Diersch (2014) and will not be
recalled here. A semi-implicit time discretization scheme with a maximum time step of 3 years advances
the coupled equations over 1 million year (Myr) for both 2D and 3D models. 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.

2.1 2D / 3D domains and faults

For the scope of this study, the LYG and its major hydrogeological features served as base for building
2D and 3D numerical models. The geometry of the two-dimensional domain illustrated in Figure 1 is
adapted from the conceptual model given in Magri et al. (2015). The profile extends over 12 km and a
500 m thick closing unit, labeled “basement”, closes the model at 6.4 km depth (Fig. 2a). This additional
layer serves as thermal buffer between bottom heat boundary conditions and the overlying 2.5 km unit
of Triassic carbonate and marly rocks (u5), which structure is to date poorly constrained. The cross-
section includes a 0.5 km thick Jurassic unit made of partly karstified limestones and shales (u4). Above,
Cretaceous rocks form the lower aquifer (u3), primarily composed of sand-, lime- and dolostone with
interbedded marls. Early Tertiary marls and chalk compose the u2 aquiclude, which form the LYG floor
and the base of its shoulders. The latter are topped by a ca. 200 m thick Tertiary limestone aquifer (u1),
which is partly covered by conductive basalts. The gorge’s shoulders quickly reach elevations of 200 m to
400 m mean sea level (MSL) and continue into heights of 800-1000 m MSL elevation. The 2D models by
Roded et al. (2013) and Magri et al. (2015) show that the measured log temperatures and the increased
geothermal gradient of the LYG are better fitted with the presence of faults or areas with enhanced
vertical permeability. Here three normal faults represented as equivalent porous media (EPM) of 40 m
width cut the aquifer system in its entire thickness.

Figure 2. Stratigraphy of the studied profile, including head level (blue line) used as flow boundary condition. The profile is
adapted from the LYG example described in Magri et al. (2015) and constitutes a schematic approach. A zoom of the finite
element mesh in the xz plane shows the grid refinement in the faults. The latter are modeled as 40 m wide Equivalent Porous
Media (EPM).

The “Triangle” algorithm (Shewchuk, 1996) is used to discretize the xz domain in finite elements with
variable width (Fig. 2, 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. Element spacing grows gradually from
the fault flanks to 50 meters in the surrounding units and basement.
The 3D solution domain is obtained by extruding the 2D mesh over 7 km in the $y$-direction (Fig. 3a). In total, the 3D model comprises 2.4 million prismatic elements with a constant resolution of 100 m in the $y$-direction (Fig. 3a). Accordingly, the faults are 40 m thick, 7 km long and approximately 5.5 km deep EPM block units. The 3D faults account for the relative displacement of the surrounding aquifers (Fig. 3b). The resulting 3D structural model is not reproducing the real structural characteristics of the LYG and is solely used as a conceptual model to investigate the evolution of convective modes in 3D.

The implementation of faults as EPM is preferred to planar or 2D discrete elements because it allows to:

1. study topography- and thermally- driven flow interactions within cutting permeable units (e.g. helicoidal flow, Fig. 3b); ii) implicitly solve conductive lateral heat exchange between fault walls and adjacent aquifers that trigger thermal convection iii) apply the theory from Wang et al. (1987), Malkovsky and Pek (2004) to determine the onset of thermal convection within the faults. This latter process is henceforth referred to as fault-convection in the fault plane $yz$ (Fig. 3b).

**Figure 3. a)** The 3D mesh is made of 2,341,850 triangular prisms of 100 m length ($y$-direction) and variable resolution in the $xz$ plane (see figure 1, $xz$ cross-section and zoom). **b)** Schematic representation of the modeled 3D faults. Faults are 40 m thick, 7 km long and ~5.5 km deep EPM units. The $yz$ surface cutting the fault is referred to as fault plane. Three different types of flow patterns involving the fault are illustrated: i) advective flow that is not affected by thermal-convection (blue line), as by example shallow or confined regional flow flowing through the fault; ii) fault-convection, i.e. convective flow developing in the fault plane, parallel to the $yz$ plane (orange dashed lines) and iii) spiral-like flow indicating the interaction between convective and advective flows. In both figures, the unit numbering ($u_1$ to $u_5$) refers to the stratigraphy given in Figure 1 and Table 1.
2.2 Physical properties of rocks and fluids

The physical properties of each unit (e.g. hydraulic conductivity, porosity and heat conductivity) are given in Table 1. The assigned values are large-scale averages derived from previous hydrogeological investigations, field studies and lithological descriptions as detailed in Magri et al. (2015). The 2D and 3D models differentiate the main aquifers and aquitards in the vertical direction (z). The anisotropy ratio of host rocks (Kz/Kx) is 0.015 while faults are isotropic. For each simulation, three faults are given identical hydraulic conductivity, which values vary between \(2.3 \times 10^{-7} \text{ m/s}\) and \(9.3 \times 10^{-7} \text{ m/s}\) (i.e. 7 m/yr and 30 m/yr, respectively). The faults and the upper aquifer have hydraulic conductivity values of the same order of magnitude, whereas the Lower aquifer (u3) can be up to 10 times more permeable than cutting units. In other words, faults do neither enhance nor prevent groundwater flow in the direction perpendicular to the fault plane.

<table>
<thead>
<tr>
<th>Units Label</th>
<th>Type of unit</th>
<th>Hydraulic conductivity (K) (m/s)</th>
<th>Thermal conductivity (\lambda) (W/m°C)</th>
<th>Porosity (\phi)</th>
</tr>
</thead>
<tbody>
<tr>
<td>u1</td>
<td>Upper aquifer</td>
<td>(2.7 \times 10^{-7})</td>
<td>2.10</td>
<td>0.06</td>
</tr>
<tr>
<td>u2</td>
<td>Aquitard</td>
<td>(3.2 \times 10^{-8})</td>
<td>1.50</td>
<td>0.08</td>
</tr>
<tr>
<td>u3</td>
<td>Lower aquifer</td>
<td>(2.1 \times 10^{-6})</td>
<td>2.24</td>
<td>0.13</td>
</tr>
<tr>
<td>u4</td>
<td>Lower aquifer</td>
<td>(7 \times 10^{-7})</td>
<td>2.50</td>
<td>0.04</td>
</tr>
<tr>
<td>u5</td>
<td>Aquiclude</td>
<td>(3 \times 10^{-10})</td>
<td>2.70</td>
<td>0.01</td>
</tr>
<tr>
<td>(F)</td>
<td>EPM</td>
<td>(9.3 \times 10^{-7}, 4.6 \times 10^{-7}, 2.3 \times 10^{-7})</td>
<td>1.1</td>
<td>0.2</td>
</tr>
<tr>
<td></td>
<td>Basement</td>
<td>(1.2 \times 10^{-12})</td>
<td>2.50</td>
<td>0.01</td>
</tr>
</tbody>
</table>

Table 1. Physical properties of the modeled units. Values are large-scale averages derived from different hydrogeological studies as listed in Magri et al. (2015). Read \(1.2 \times 10^{-12}\) as \(1.2 \times 10^{-12}\). Longitudinal and transversal dispersivity are null, i.e. the tensor of thermodispersion is \(\Lambda = \phi \lambda^l + (1 - \phi) \lambda^t\).

Fluids properties are summarized in Table 2. The polynomial fittings of Magri et al. (2015) are used to compute variable fluid viscosity \(\mu^f\) and density \(\rho^f\) over the pressure and temperature ranges of the
solution domain. No concentration effects are accounted because of the low salinity that characterize the western LYG groundwater. Fluid thermal conductivity $\lambda^f$ and heat capacity $c^f$ are constants at reference temperature of 20 °C.

<table>
<thead>
<tr>
<th>Fluid property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Thermal conductivity $\lambda^f$ (W/m/°C)</td>
<td>0.65</td>
</tr>
<tr>
<td>Volumetric heat capacity $\rho^f c^f$ (J/m$^3$/°C)</td>
<td>4.2e6</td>
</tr>
<tr>
<td>Fluid density $\rho^f$ (kg/m$^3$)</td>
<td>Variable</td>
</tr>
<tr>
<td>Fluid viscosity $\mu^f$ (Pa/s)</td>
<td>Variable</td>
</tr>
<tr>
<td>Coefficient of thermal expansion $\beta$ (°C$^{-1}$)</td>
<td>Variable</td>
</tr>
</tbody>
</table>

Table 2 Fluid properties. Variable fluid properties are pressure and temperature dependent expressions computed with the polynomial fittings given in Magri et al. (2009). Read 4.2e6 as 4 x 10$^6$ (computed with an Equation of State). For the Rayleigh calculations the used values are those at reference temperature and pressure (i.e. minima values). No salinity effects are accounted in the simulations.

2.3 Boundary and initial conditions

2D and 3D models use identical boundary conditions, as follows: The groundwater levels vary from -200 m MSL at the left ending of the domain to 50 m MSL at the right ending. Though these piezometric measurements are characteristics of the LYG, they induce a steady state regional flow over the 12 km domain extension that can be encountered in any basin system, where highlands and lowlands act respectively as recharge and discharge areas. When a hydraulic gradient is imposed to a thermal buoyant flow, the resulting process is called mixed convection. By contrast, in a free thermal regime, no regional flow exists. For the specific case of free thermal convection, a flat head boundary condition is set at the top of the domain. No flux boundary conditions are assigned on the lateral sides and bottom of the domain. The temperature at the top of the domain is allowed to vary with respect to convective heat transfer through a Cauchy boundary condition. The reference temperature is 20 °C.
constant heat flow of 60 mW/m² is imposed at the basement. Lateral boundaries are thermally insulated.

The initial conditions for head and temperature distributions are derived from steady state simulations of groundwater flow and conductive heat transport, respectively.

3. Results and discussions

3.1 2D results

Linear stability analysis based on Rayleigh number calculations offers a useful tool to determine the onset of thermal convection (e.g. Nield, 2006). The Rayleigh number is:

\[ R = \frac{K d \beta \Delta T}{\lambda} \]  

(1)

where \( K \) (m/s) is the hydraulic conductivity of the aquifer, \( \beta \) (1/°C) is the coefficient of thermal expansion and \( \rho_f \) (kg/m³) the fluid density. \( \Delta T \) (°C) is the variation of temperature over the aquifer thickness \( d \) (m). \( D = \frac{\phi \lambda_f + (1-\phi) \lambda_s}{\rho_f c_f} \) (m²/s) is the tensor of thermodispersion where \( c_f \) (J/kg/°C) is the specific heat capacity of the fluid, \( \lambda_f \) and \( \lambda_s \) (W/m/°C) are the thermal conductivity of fluid and solid, respectively.

Considering units u3 and u4 define together a 2 km aquifer (Fig. 2, Fig. 3) with thickness-weighted average values for porosity, permeability and heat conductivity of 0.1, 1.4e⁻⁶ m/s and 2.4 W/m/°C and assuming that \( \Delta T \approx 60 \) °C over the 2 km thick aquifer and estimating that therein \( \beta \approx 3e^{-4} °C^{-1} \), then:

\[ R \approx \frac{1.4e^{-6} \times 2000 \times 3e^{-4} \times 60}{(0.1 \times 0.65 + (1-0.1) \times 2.4) / 4.2e^6} = 95 \]
which is higher than $4\pi^2$, the critical Rayleigh number $R_c$ for an infinitely long homogenous porous media (Lapwood, 1948). This analytical estimation suggests that thermal convection likely develops within the lower aquifers, as shown by the numerical simulations. 2D results of coupled fluid flow and heat transport processes are illustrated for both mixed and free convection in Figure 4 for the case $K_{\text{fault}} = 9.3 e^{-7}$ (m/s), i.e. 30 m/yr.

Figure 4. 2D results. a-b) Darcy velocity (m/yr), and c-d) temperature (°C). Mixed convection regime (a-c) occurs when topography- and thermally-driven flow interacts (white and red pathlines, respectively). Free convection (b-d) refers to a purely buoyant-driven flow simulated with a flat head boundary condition at the top, preventing any imposed regional flow.

In both cases, two convective cells develop in the lower aquifers u3 and u4 at each side of the fault F3 (Fig. 4a; b). The cells are layered and bounded at the top and bottom by the aquitards u2 and u5, respectively. Convection circulates deep groundwater at peak velocity of 0.2 m/yr. The flow boundary condition at the top controls the direction of convective circulation: In the mixed convection regime (Fig. 4a), the cells drive thermal water upward in the “gorge fault” F3, whereas in the free regime (Fig. 4b) the flow along the F3 fault is downward. As a consequence, the number of protruding thermal plumes
differs in each scenario (Figs. 4c, d). The mixed convective regime generates two thermal plumes reaching the surface (Fig. 4c). In this case, the simulated spring temperatures range between 40 °C and 50°C. In the free thermal regime (Fig. 4d), the temperature gradient below the gorge is decreased. Only the discharging fault F1 at the left ending of the profile flushes thermal water at temperature close to 60°C. In both mixed and free convective regimes, no thermal convection develops within the faults. These EPM units act as preferential pathways for either ascending or descending groundwater. In 2D only one flow direction is possible within the fault.

The described patterns are sufficient to highlight the general mechanisms of coupled fluid transport processes, namely topography-driven flow (regional flow) superimposed to thermal convection that occurs in many faulted hydrothermal systems. The resulting mixed convective regime can explain flow directions along faults and related increased/decreased temperature gradients. However, because of their 2D constraints, the models cannot capture modes of convection that are parallel to the fault plane (i.e. perpendicular to the described profile) thereby impeding any comparison between the spatial distribution of convective cells and the occurrences of thermal springs along the fault trace. The next 3D simulations are aimed to study the interaction between fault convection (Fig. 3b, orange cells) and crosscutting groundwater flow, which can be either regional flow (Fig. 3b blue vector) or thermal convective flow in the surrounding aquifers (Fig 3b, red vectors).

### 3.2 3D Results

Wang et al. (1987) demonstrated that the critical Rayleigh number $R_c$ of 3D vertical faults embedded in impervious rocks is inversely proportional to the fault aspect ratio (fault width/height), i.e. referring to Figure 2b, $R_c \propto O(H/d)$. Malkovsky and Pek (2004) provided an analytical relationship to calculate $R_c$ values, for the case of a fault with a permeable top, as follows:
$$R_c = \left( \frac{6.428}{\Delta} \right)^{1.165} + (27.1)^{1.165} - 0.8584^d, \text{ with } \Delta = \frac{d}{2H} < 0.1 \text{ (half of the aspect ratio)}$$  

Tournier et al. (2000) calculated similar values of critical Rayleigh numbers by means of numerical simulations. Furthermore, their numerical benchmarks showed that at those $R_c$, thermal convection in the fault is triggered after a certain time-period and therefore is also referred to as delayed convection. A limitation of the mentioned studies is that fluid viscosity is constant, although small viscosity changes can exert a substantial influence on conditions for onset of free thermal convection (Zhao et al., 2006).

Recent analytical and numerical investigations by Malkovsky and Magri (2016) estimated that in a faulted geothermal system under normal temperature conditions as those modeled here, variable fluid viscosity triggers thermal convection within faults which permeability is four times lower than in the case of a fluid with constant viscosity. In this study, since fluid viscosity is temperature dependent, it is assumed that fault convection occurs at $R_c / 4$ where $R_c$ is estimated at lowest temperature conditions.

In the studied case (Fig. 3b), $\Delta \cong 40 / (2 \times 5500) = 3.6e^{-3}$ (half aspect ratio) leading to the critical Rayleigh number (Eq. 2) $R_c \approx 1780 / 4 = 445$ for a temperature-dependent viscous fluid.

The critical hydraulic conductivity value of the fault, $K_{crit}$, that corresponds to $R_c = 445$ can be derived from Eq. 1, using the fault and fluid properties given in Table 1 and Table 2, respectively. Estimating that over the 5500m deep fault $\Delta T \cong 140$ °C, a thermal expansion coefficient $\beta \approx 5.9.e^{-4}$ °C$^{-1}$ provides a good fluid density fit (e.g. Fig. 4 in Malkovsky and Magri, 2016) then:

$$K_{crit} \approx \frac{445 \times (0.2 \times 0.65 + (1 - 0.2) \times 1.1) / 4.2e^6}{5500 \times 5.9e^{-4} \times 140} \approx 2.4e^{-4} \text{ m/s}$$

i.e. fault convection is favored at $K_{\text{fault}} = 9.3e^{-7}$ m/s and $K_{\text{fault}} = 4.6e^{-7}$ m/s (Table 1). At $K_{\text{fault}} < 2.4e^{-7}$ the thermal regime is purely conductive.
In narrow apertures, the fault plane $yz$ is the preferred mode of convection (Simmons et al., 2008), and consists of weakly three-dimensional square cells (Wang et al., 1987), as illustrated by orange cells in Figure 2b. The next 3D simulations will provide insights into the interactions between fault convection and lateral cross-cutting regional flow.

Results of 3D mixed convective flow are illustrated in Figure 5 for the case $K_{\text{fault}} = 9.3 \times 10^{-7}$ m/s (i.e. 30 m/yr). The cross-sectional views parallel to the regional flow (Figs. 5a; b) reveal that fluid convection can develop within the major aquifers u3 and u4. The location and shape of the cells however differ from the 2D case (compare with Figure 4a). By example, in the middle slice (Fig. 5b), groundwater flows from the right ending of the profile directly toward the F3 fault, whereas in 2D the aquifers are characterized by a persistent convective regime adjacent to the faults and a bounded regional flow in the top u1 aquifer (Fig. 4a). In 3D, multi-cellular motion is dominant in each fault plane (Fig 5c), as predicted by analytical calculations. Within the faults, two elongated convective cells stretch over the entire fault height. The cellular motion consists of a central downflow and two lateral upflows at velocities ranging between 0.5 m/yr and 2 m/yr, respectively. Therefore, inflow and outflow of groundwater can occur within the same fault over its entire length (i.e. y-direction). Fault convection allows groundwater to reach basement depths where it gets heated before starting its upward migration within the fault. During this process, two scenarios are then possible: i) thermal water discharges through the fault surface mixing with the regional flow (red and white arrows in Fig. 5c) ii) ascending thermal water interacts with the advective regional flow that enters the fault through the intersecting u3 and u4 aquifers (yellow circles in Figure 5c). In the latter case, the direction of the regional flow outside the fault is deviated or undergoes complex helicoidal (spiral-like) patterns (red circles in Fig. 5c), as illustrated in the 3D rendering (Fig. 5d).
Figure 5. Mixed convection. $K_{\text{fault}} = 9.3 e^{-7}$ m/s (i.e. 30 m/yr). Flow paths plotted over the velocity field (meter per year) for different views. Cross-sectional xz views (a - b) show and the upper regional flow (white lines and vectors) as well as different cross-sectional convective patterns (black, blue, green and red lines) in the aquifers u3 and u4. Fault plane yz views (c) indicate that two convective cells develop in each fault (red lines, vectors). Groundwater entering/leaving the fault perpendicular to the fault plane (i.e. in the x direction) is schematically represented with a red/yellow circle, as also explained in figure 2b. White arrows depict regional flow recharge/discharge areas (d). The black, blue, green and red flow paths as shown in a) - b) are also plotted in a 3D top view revealing complex helicoidal flow patterns. As a result, thermal water can discharge at different locations forming cluster of springs close to fault traces.

The coexistence of these flow processes, namely regional flow in the aquifers and convection within the fault, strongly impacts the temperature field (Figs. 6a; b). Temperature decreases in the center of the fault, in relation to downward flow and increases along the fault sides (Fig. 6c). Furthermore, the spacing between thermal plumes is controlled by cellular motion in the faults. Accordingly, the temperature profiles vary also along the y-direction (i.e. xz cross-sections perpendicular to the fault plane, Fig. 5b), in contrast to the purely two-dimensional case (Fig. 4c). As the regional flow determines the location of the discharge areas, springs form at the top of F1 and F3 faults. Along those fault traces, the locations of springs correspond to the wavelength of the underlying convective cell. However, since the convection is highly transient in nature, a given spring might shift along the fault trace, depending on the periodicity of the cellular motion (Fig. 6d). Springs temperature can reach 60°C. An additional simulation in which $K_{\text{fault}} = 4.6 e^{-7}$ m/s (i.e. 14.5 m/yr) computed qualitatively very similar flow patterns. The only observed
differences are lower convective velocities and less decreased temperature field in the central part of the faults.

Figure 6. Mixed convection. $K_{\text{fault}} = 9.3 \times 10^{-7} \text{ m/s}$ (i.e. 30 m/yr). Calculated temperature (°C). Thermal plumes are illustrated along different cross-sectional xz views (a - b). Fault plane yz views (c) show the effects of convection: temperature decreases in the central part of the fault and increase at both lateral sides. The 3D top views (d) at 600,000 years and 1,000,000 years indicate that thermal springs can shift over time along fault traces because of the transient nature of fault convection. Zooms of the clusters (black rectangles in (d)) show that the temperature difference between closely interspaced springs can be more than 10°C. Simulations with $K_{\text{fault}} = 4.6 \times 10^{-7} \text{ m/s}$ (i.e. 14.5 m/yr) display very similar general features.

In the next simulation, the hydraulic conductivity of the fault is $K_{\text{fault}} = 2.3 \times 10^{-7} \text{ m/s}$ (i.e. 7.2 m/yr). Here it is worth recalling that the faults are 3 to 10 times less permeable than the aquifers u4 and u3, respectively (Table 1). However, the relatively small thickness of the faults with respect to the km-scale extension of the study domain does not prevent advection across the faults. A striking result is the onset of fault convection despite the simulated hydraulic conductivity of this scenario falls below the previously estimated critical value of $2.4 \times 10^{-7} \text{ m/s}$. Compared to the previous cases ($K_{\text{fault}} = 9.3 \times 10^{-7} \text{ m/s}$ and $K_{\text{fault}} = 4.6 \times 10^{-7} \text{ m/s}$), the thermal plumes in the F2 and F3 faults are smoothed (Fig. 7a) suggesting that therein the regional flow does overwhelm thermally-driven flow. However, the convective and helicoidal 3D flow paths (Fig. 7b) close to the F1 fault indicate that buoyant and thermal forces still interact within discharge areas. The fault plane views of the velocity field (Fig. 7c) further clarifies the ongoing
processes: groundwater flows into the upper 3 km of the faults through the permeable aquifers u3 and u4 (Fig. 7c, blue circles) at velocities between 0.1 m/yr and 3 m/yr whereas, at Triassic depths, a single elongated convective cell drives thermal water in the fault plane at maximum velocity of 0.2 m/yr. The thermal water exits the F3 and F2 faults along the top Triassic (Fig. 7c, red circles) to eventually discharge at the surface of the F1 fault where temperatures can locally be higher than 40 °C (Figs. 7a; d).

Without the regional flow (free convection), the simulated thermal regime remains purely conductive, in agreement with the analytical calculations previously described. Therefore, the presence of deep-seated convective cells within less permeable faults is due to the crosscutting regional flow that transports additional lateral heat to the fault walls, thereby triggering buoyancy.
Figure 7. Mixed convection. $K_{\text{per}} = 2.3 \times 10^{-7}$ m/s (7.25 m/yr). (a) 3D front view of the calculated temperature (°C). Within a predicted cluster of springs, temperature varies of 5°C (b) 3D flow paths show both fault convection and helicoidal features. (c) Fault plane yz views of the velocity field (m/yr) indicate that one convective cells develop in each fault, at Triassic depth (red lines, vectors). By contrast, the regional flow is dominant in the upper fault entering F2 and F3 through intersecting aquifers (blue circles). Thermal water enters/leaves the fault along the top permeable aquifer u4. (yellow/red circles). Groundwater captured by fault convection in F1 (yellow circles) discharges at the surface mixing with regional flow (white arrows). The resulting temperature field (d) is illustrated at each fault plane yz views.

In their pioneering simulations of hydrothermal circulation in fault zones, Lopez and Smith (1995) mapped permeability ranges that delimit the transition from convective to advective regimes. They found that in a permeable fault zone, convection is favored over advection when the permeability of the country rock is lower than given values, in agreement with the numerical observations presented here. In this specific scenario, the estimated hydraulic conductivity of the hosting rock that favors convection in the fault is $K \leq 7 \times 10^{-7}$ m/s (Table 1). While Lopez and Smith (1995) consider a fault cutting a single...
aquifer, here it is suggested that in a system characterized by multilayered aquifers, different transport regimes can coexist within a fault zone as the result of the interplay between pressure forces and buoyancy. In this respect, permeability distributions of the surrounding aquifers and fluid density variations play a key role as they respectively control advective fluxes and buoyant forces.

A final simulation of free convective flow is carried out for the case $K_{\text{fault}} = 4.6 \times 10^{-7}$ m/s (14.5 m/yr), i.e. at conditions favoring fault convection. Results are illustrated in Figure 8. It can be seen that patterns and values of both velocity and temperature fields (Fig. 8a; b) are very close to the two-dimensional results shown in Figures 4b and 4d. In 3D, the anti-clockwise convective cell at the right side of the F3 fault extends over the y-direction forming elongated rolls. At the intersection with fault cells, these rolls assume spiral-like shapes, as it can be seen in Figure 8c, close to the F1 and F2 faults. The mechanism is analogous to the previously explained mixed convective case, where the interactions between thermal convection in the fault plane and the imposed transversal regional flow led to helicoidal flow patterns.

![Figure 8. Free convection. $K_{\text{fault}} = 4.6 \times 10^{-7}$ m/s (i.e. 14.5 m/yr). (a) Flow paths plotted over the velocity field (m/yr) for a cross-sectional xz view (a) 3D front view of the calculated temperature (°C). (b) 3D flow paths show convective rolls (blue cells) and fault convection. The interaction between these two processes generates spiral-shaped flow patterns (c).](image-url)
Conclusions

Though based on simplified structural hydrogeological features, the simulations presented here provided insights into the possible interactions between deep flow processes in faulted basins that could not be inferred from 2D studies. Specifically, the sensitivity analysis of the fault permeability showed that faults crosscutting the main regional flow direction allow groundwater to be driven laterally by convective forces within the fault planes (Figs. 5 and 7). Therein thermal waters can either discharge along the fault traces or exit the fault through adjacent permeable aquifers. In the latter case, the resulting flow is helicoidal and transient (Figs. 5d; 7c; 7d). Accordingly, the location of discharge areas can move with time and is not strictly constrained to the damage zones (Fig. 6d). The spacing between cluster of springs reflects the wavelength of the multicellular regime in the fault zone. Vigorous regional flow favors deep layered convection within fault planes by enhancing lateral heat transfer to the fault walls. Simulations show that both regimes can coexist also at values of fault hydraulic conductivity below critical Rayleigh conditions (Fig. 7). By contrast, when the regional flow is weak or totally absent, as in free convection (Fig. 8), fault convection is triggered only above Rayleigh conditions, as predicted by the viscous-dependent analytical solutions from Malkovsky and Magri (2016). Furthermore, in a free convective regime, convective rolls similar to those observed in the 2D case (Figs. 4b; d) extend throughout the entire aquifer length (Fig. 8c). The interactions between these 2D-like convective rolls and planar fault cells generate spiral-like transient patterns, as those observed in the mixed convective regime.

The idealized numerical models presented here suggested possible transport mechanisms that can, for example, be applied to the LYG (Fig. 1) where thermal water ascends through fractured aquicludes and the hydrochemical character of the emerging groundwaters is disparate. Investigations on rare earth elements, major elements and stable isotopes show that along the level course of the LYG and below the Golan Heights, NE-SW and N-S oriented deep flow paths must exist to permit the chemical
composition of the groundwaters in and along the LYG, respectively (Siebert et al., 2014). The idealized numerical models presented here support the hypothesis of crossing flow paths, resulting from the coexistence of fault convection, that can develop for example along longitudinal East-West faults within the Gorge, and additional flow fields that can be induced by topography gradients, e.g. along N-S axe of the Gorge, or by local thermal convection in permeable aquifers below Eocene aquiclude. Given the current estimation of damage zones width (40 m), fault height (5.5 km), and basal heat flow of 60 mW/m$^2$, a fault hydraulic conductivity of $2.4 \times 10^{-7}$ m/s ($7.2$ m/yr) would be sufficient to induce thermal flow in fractured Triassic sediments. Theoretically, deep fluids can migrate through the entire fault length before being flushed in discharge areas by topography-driven flow. At hydraulic conductivity higher than $4.3 \times 10^{-7}$ m/s ($13.6$ m/yr), fault convection appears to be controlling heat fluxes also at shallower depths. This highly transient process likely underlies the formation of clustered thermal springs along the Lakeshore. As it is recognized that thermal convection can be a driving mechanism for migration of hydrocarbons (Rabinowicz et al., 1985), the complex helicoidal patterns described here could support the highly debated hypothesis of oil accumulation under sealing caprocks.

At the current state, further investigations are being carried out to better constrain the structural features of the basin, such as orientation, depth of penetration and intersections of the faults. For example, numerical calculations from Person et al. (2012) suggest that at given permeability conditions fault intersections focus outflow of hot groundwater. Further aspects that need to be addressed are the effects of aquifer and fault heterogeneities (e.g. Ingebritsen and Gleeson, 2014; Lopez and Smith, 1996) and mineral precipitations that can potentially clog faults over the geological time-scale considered here, thereby hindering convective flows (e.g. Genthon et al., 1997).

Despite these model simplications and assumpitons, it has been shown that increases of surface temperature near faulted areas fall within measured ranges ($20^\circ$C - $60^\circ$C), suggesting that co-existing modes of thermal convection are likely controlling transboundary flow across the LYG:
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