NUMERICAL ANALYSIS OF HEAT TRANSPORT WITHIN FRACTURED SEDIMENTARY ROCK: IMPLICATIONS FOR TEMPERATURE PROBES

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Abstract:
Thermal energy transport from ground surface into a fractured sedimentary rock aquifer is simulated numerically to gain insight into the response of temperature probes for identifying hydraulically active fractures in rock boreholes. The conceptual hydrogeological model is based on a field study site at Guelph, Ontario, Canada, where thin sandy overburden overlies a densely-fractured dolostone. Characterizing the fracture network is important in this area for understanding the behaviour of DNAPL contaminants at nearby industrial sites. The model includes density-dependent groundwater flow coupled with thermal advection, conduction, and retardation within the porous matrix and discrete fractures. The fracture elements are generated stochastically and are coupled directly with the porous medium matrix blocks. Natural background flow gradients, surface recharge and seasonally variable surface temperatures are also considered. The results show that ground source thermal pulses can propagate deep into a fractured rock system and appear as weak thermal “anomalies” within the fractures on the order of a few tenths or hundredths of degrees. Such anomalies are now becoming detectable with state-of-the-art thermal probe technologies and can be potentially valuable natural signals for identifying hydraulically active fracture zones.
Introduction

Predicting the behaviour of groundwater contaminants through fractured porous media remains one of the most challenging fields of hydrogeology (MacQuarrie & Mayer, 2005; Neuman, 2005). Much of the uncertainty originates from the unknown fracture geometry including fracture length, spacing, orientation and aperture distributions. Further uncertainty arises from the flux distribution within these networks since not all fractures are conductive.

Well-known for their use in porous media, geophysical temperature logs within fractured rock environments have recently re-emerged as a promising technique for characterizing active fracture networks. In particular, continuous temperature logging within lined boreholes has shown significant advantages over the more conventional open-hole logs (Greenhouse & Pehme, 2002; Pehme et al. 2007). Numerical modelling can be used to help understand the response of these highly sensitive temperature probes.

In this paper, we present conceptual numerical simulations of heat transport within discretely fractured porous sedimentary rock. Natural heat energy pulses are generated using seasonal air temperatures applied as a thermal flux condition uniformly over the upper boundary surface. The objective is to determine if natural temperature signals can propagate from ground surface deep within fractured rock. The results will be useful to help interpret signals detected using advanced temperature probes, such as described by Pehme et al. (2007).

Numerical Simulation Approach

The three-dimensional (3D) discrete fracture network (DFN) model SMOKER (Molson & Frind, 1994, 2005) is used in this study to simulate thermal transport (the name derives from the Black Smoker hydrothermal systems to which the model was applied by Yang et al. 1996). Coupled groundwater flow and thermal transport is considered in both the porous medium and fracture network. The heat transport equation for the porous matrix can be expressed as (Molson et al. 1992):

$$\frac{\partial}{\partial x_i} \left[ \left( \kappa + \frac{D_y}{R} \right) \frac{\partial T}{\partial x_j} \right] - \frac{\partial}{\partial x_i} \left( \frac{\nu_i}{R} T \right) = \frac{\partial T}{\partial t}$$

(1)

where $T$ is the temperature, $D_y$ is the hydrodynamic dispersion tensor ($m^2 s^{-1}$), $\nu_i$ is the average linear groundwater velocity ($m s^{-1}$) (determined from the hydraulic head solution) and $\kappa$ is the aquifer thermal diffusivity ($m^2 s^{-1}$) defined by $\kappa = \lambda / C_o$ where $\lambda$ is the aquifer thermal conductivity ($J m^{-1} s^{-1} \circ C^{-1}$) and $C_o$ is the aquifer heat capacity ($J m^{-3} \circ C^{-1}$).

The thermal retardation $R$ for the porous matrix is defined by

$$R = \frac{C_o}{S \cdot \Theta \cdot c_w \cdot \rho_w}$$

(2)
where $\theta$ is the porosity, $S$ is the degree of water saturation, $c_w$ and $\rho_w$ are the specific heat (Jkg$^{-1}$ °C$^{-1}$) and density of water (kgm$^{-3}$), respectively. Neglecting the air phase, the heat capacity of a porous medium is given by

$$C_o = S\theta c_w \rho_w + (1-\theta)c_s \rho_s$$

(3)

where $c_s$ is the specific heat of the solid phase (grains or rock) (Jkg$^{-1}$ °C$^{-1}$) and $\rho_s$ is the solid phase density (kgm$^{-3}$).

The governing equation for thermal transport within a planar fracture can be written as:

$$\frac{\partial T'}{\partial t} + \frac{\partial (\bar{v}_i T')}{\partial x_i} - \frac{\partial}{\partial x_i} \left( \frac{\lambda_w}{c_w \rho_w} \frac{\partial T'}{\partial x_i} \right) - \frac{\lambda_m}{c_w \rho_w b} \left[ \frac{\partial T'}{\partial z} \right]_{z=\pm b} = 0 \quad (i, j = x, y)$$

(4)

where $T'$ is the fracture temperature (°C), $\bar{v}_i$ are the fracture velocities (ms$^{-1}$), $\lambda_w$ and $\lambda_m$ are the respective thermal conductivities of water and the porous matrix (Jm$^{-1}$s$^{-1}$°C$^{-1}$), and $b$ is the half-fracture aperture (m). The last term on the left hand side of (4) represents the heat transfer between the fracture and the porous matrix.

The fracture velocities for equation (4) are determined in the model using the simulated hydraulic head field and using the cubic law which can be written as:

$$\bar{v}_i = \frac{-(2b)^2}{12\mu} \rho g \nabla h$$

(5)

where $2b$ is the full fracture aperture (m), $\mu$ and $\rho$ are the viscosity ($10^{-3}$ kgm$^{-1}$s$^{-1}$) and density (kgm$^{-3}$) of water, respectively, $g$ is the gravitational acceleration (ms$^{-2}$) and $\nabla h$ is the hydraulic head gradient (-). For example, assuming a gradient $\nabla h=0.01$, $2b=1000$ µm ($10^{-3}$m), $\rho=1000$ kg/m$^3$, $\mu=10^{-3}$ Pa s, then $v = 0.0083$ m/s, or ~720 m/d. Note that thermal conduction from the fracture into the matrix will retard the migration velocity of the thermal pulse through the fractured rock.

Boundary conditions for equation (1) can be the standard first, second or third-type (Molson and Frind, 2005). The equations are solved using a finite element approach with rectangular prism elements for the porous matrix and planar elements for the fractures. The fracture elements are embedded directly into the porous medium and equations (1) and (4) are solved simultaneously allowing for advective-dispersive transport in both the porous medium and matrix blocks. Fluid density and viscosity are temperature-dependent.
2D Single Fracture System

Heat Transport

The behaviour of heat transport within fractured sedimentary rock is first studied by applying the SMOKER model to a single fracture within a porous matrix; the analysis includes a comparison against the recently-developed SFRAC-H analytical model (Meyer, 2004).

The conceptual system for the single fracture case is shown in Figure 1. Flow in the matrix is assumed negligible (a limitation of the analytical solution), and the velocity in the horizontal fracture is 0.005 m/s. Except for the fixed temperature at the fracture inflow boundary node (type-1 Dirichlet condition), all transport boundaries are type-2 (Neumann) zero-gradient condition. We assume a fracture aperture (2b) of 1mm, a thermal conductivity of 2.0 J/m/s/°C, a solids heat capacity of 860 J/kg/°C, and a solids density of 2630 kg/m³.

After 3 days, the 11.5 °C temperature front has advanced about 6m, and has extended about 1.2m into the matrix at the inflow boundary (Figure 1). The longitudinal thermal profiles within the fracture are shown in Figure 2a. The fit of the numerical model to the analytical solution shows that the physical processes have been accurately reproduced.

![Figure 1](image-url)  
Figure 1. Conceptual layout and model parameters used for model validation. Two-dimensional, single planar fracture system, numerical temperature solution shown at 3 days.
Figure 2. Model validation and heat vs. mass transport comparison in a single fracture system (see Fig. 1) showing (a) comparison of the numerical model (SMOKER) against the analytical model (SFRAC-H), and (b) simulated concentration profiles from an otherwise equivalent mass transport simulation using SMOKER and CRAFLUSH (Sudicky, 1988) (for clarity, the numerical solutions are only plotted every 3 points).

Heat vs. Mass Transport

To highlight the difference between heat transport and mass transport within a fractured porous system, an otherwise equivalent mass transport simulation is repeated for the above system (Figure 1), assuming a fixed concentration of Co=15 at the inflow boundary and an initial concentration C₀=11. In this case, thermal diffusivity is replaced by hydrodynamic dispersion, assuming a longitudinal dispersivity in the fracture of αₐ=0.1m, and a diffusion coefficient of 10⁻⁹ m²/s. The longitudinal concentration profiles are provided in Figure 2b, at travel times where the leading edge concentrations match the leading edge temperatures from the profiles in Figure 2a.

Two trends are immediately clear on comparison of Figure 2a and 2b: first, a significant retardation is induced by thermal conduction from the fracture to the porous matrix (note profile times), and second, while the concentration profiles are typical of advective-dispersive transport, the thermal profiles are significantly depressed (almost diffusive-like), which is again due to thermal loss to the matrix. In comparison, the concentration profiles show limited diffusive loss to the matrix at these early times. Comparing the travel times for the corresponding temperature and concentration profiles, effective thermal retardations of approximately 80, 140, 214 and 500 can be derived for the thermal profiles at 0.25, 1, 3 and
10 days, respectively. The thermal retardation grows in time since the thermal system between the fracture and matrix remains in disequilibrium. The retardation will depend on the thermal properties of the medium, and on the fracture spacing. In a system with closely spaced fractures which are all carrying a thermal plume, the matrix will reach thermal equilibrium more rapidly and thus thermal retardation will be less. In the extreme case where the fracture density approaches the grain scale, thermal retardations would approach 2-3, which is typical of a sandy aquifer (Markle et al. 2006).

Heat loss to the matrix of a fractured rock is significantly higher than mass loss because the matrix thermal diffusivity \( (\kappa) \) (here, \( \kappa = 8.8 \times 10^{-7} \text{ m}^2/\text{s} \)) can be up to three orders of magnitude higher than a typical aqueous diffusion coefficient. For the same reason, heat stored within the matrix will diffuse rapidly back to an active fracture if the temperature gradient reverses. In the following sections, the impact of this high thermal diffusivity and thermal retardation will be shown for larger scale fractured systems.

2D Fracture Network Simulations:
Natural temperature signals in fractured rock

The SMOKER model is applied herein to simulate thermal transport within a fractured porous rock aquifer. A seasonally variable natural temperature source is assigned as the upper boundary condition and propagation of the thermal signals is simulated over time. The simulations are used to investigate thermal probe temperature signatures for characterizing fracture networks. Implications regarding use of temperature probes are discussed.

Conceptual model and boundary conditions

The conceptual model is based on observed conditions at two NAPL-contaminated field sites located near Guelph and Cambridge in southern Ontario (Turner, 2001; Burns, 2005). The simplified site model consists of a 2D vertical section domain, 200m long by 50 m deep, with 5m of sandy overburden overlying 45m of the fractured dolostone aquifers of the Guelph and Eramosa Formations (Figure 3). The upper model boundary represents ground surface. Flow is assumed negligible in the transverse horizontal dimension.

The hydraulic conductivity and porosity are assumed in the model to be \( 10^{-7}\text{m/s} \) and 0.1, respectively, for the dolostone matrix and \( 10^{-5}\text{m/s} \) and 0.3, respectively, for the sand overburden. The mean fracture aperture of the random network (Figure 4) is 500 µm and the thermal conductivity of the dolostone and sand is assumed equal to 2 J/m°C. The heat capacity and density for the dolostone are \( 800 \text{ J/kg°C} \) and \( 2630 \text{ kg/m}^3 \), respectively, while those for water are \( 4174 \text{ J/kg°C} \) and \( 1000 \text{ kg/m}^3 \). Water density and viscosity are temperature dependent (Molson et al. 1992) but changes in these properties are not significant at the natural background temperatures considered herein.

The fractured porous medium is assumed saturated. A background flow system from left to right is generated by imposing fixed heads of 51 and 50m at the left and right
boundaries, respectively, assigning a uniform recharge of 400 mm/yr across the top and assuming an impermeable lower boundary.

Figure 3. Conceptual model for the heat transport simulations, showing the fracture network and flow boundary conditions.

Figure 4. Simulated fracture aperture distributions.

For the thermal transport problem, a temperature-dependent thermal heat flux is imposed across the upper boundary, which takes the form:

$$ J_i = \left( \frac{\lambda_u}{B_z} + qc_s \rho_w \right) (T_a - T_s) $$

(5)

where $J_i$ is the thermal flux across the boundary (J/m²/s), $\lambda_u$ is the thermal conductivity of the surface layer (J/m°C), $B_z$ is the surface layer thickness (m), $T_a$ is the air temperature (°C) (given), and $T_s$ is the surface aquifer temperature (°C) (computed in the model). All other boundaries are assigned a type-2 (Neumann) zero-gradient condition. For the cases herein, we assume a sinusoidal air temperature variation with a period of 365 days, a mean of 12°C, and an amplitude of 12.5°C. The surface layer properties are assumed $\lambda_u = 1$ J/m°C, and $B_z = 0.1$m.
The simulation was run until the effect of the initial conditions had disappeared, which took on the order of 20 years. Simulation times are referenced with respect to the cycle time $t_c$, which represents the time within one cycle of 0-365 days, after quasi steady state (seasonally oscillatory) conditions had been achieved.

2D Simulation Results

The simulated temperature distribution in the 2D section is shown in Figure 5 for three cycle times. The temperature distribution at each time is characterized by increasing temperature from the bottom fixed boundary ($5^\circ$C) to within about 10m of the top surface, where the temperature fluctuates, sometimes decreasing with elevation, depending on the seasonal air temperature.

Superimposed on this general trend are numerous temperature disturbances, or anomalies, caused by the fracture network in this otherwise homogeneous rock. In Figure 5a, for example, between about 75-200m, the temperature contours have been displaced downwards, caused by a relatively dense fracture zone in this area within the lower half of the domain, which carries upgradient shallower and hence warmer water downward and to the right, passing below the relatively unfractured zone centered at about $(x,z) = (125, 25m)$. Similar trends can be seen at the other times shown in Figures 5b and 5c, including some shallower anomalies between about 30-40m elevation.

While these simulated temperature anomalies are clearly related to the fracture network, it is not immediately clear if the temperature distribution can be used to identify individual fractures. Thermal conduction into and out of the porous matrix has apparently masked much of the effect of individual fractures. In the following section, however, we will show that these temperature distributions indeed contain some very useful data on the fracture network.
Figure 5. Simulated temperature distribution within the 2D section at cycle times ($t_c$) of (a) 105 days, (b) 200 days and (c) 355 days. Insets show one annual cycle of air temperature and the corresponding cycle plot times for each section. The steady state active fracture network is visible in Figure 5a from a superposition of the fracture velocities.
Temperature Profile Analysis

Small-scale fracture-induced anomalies in the temperature simulation were identified by extracting vertical temperature profiles at two locations (x=50 and x=130m), and at two cycle times (t_c=105 and 355 days) from the 2D simulation of Figure 5. These profiles were then smoothed using a moving boxcar filter, and subtracted from the original raw profiles to obtain a temperature difference (ΔT) profile. Vertical profiles of fracture velocities (Vf) for horizontal fractures intersecting these profiles were also extracted. The results are shown in Figures 6 and 7 for cycle times of 105 and 355 days, respectively.

Temperature anomalies occur along the vertical profiles because of complex fracture interconnections which affect the age of groundwater as it reaches the profile along horizontal fractures. Since the ground surface temperatures are time-dependent, different flow paths intersecting the profiles will bring water of different temperatures. A temperature anomaly will be produced at the profile if the water temperature in the fracture is significantly different from the ambient temperature above and below the fracture.

In all profiles of Figures 6 and 7, there is an excellent correlation between the fracture velocity profiles and the corresponding temperature difference anomalies. While not every active fracture produces an anomaly, every anomaly has a clearly identifiable fracture or fracture zone associated with it. The correlation also seems to be better in the downgradient section (BB'), likely as a result of higher velocities as more recharge is captured along the downgradient direction, and as more flow is concentrated in discrete fracture zones.

In Figure 6b, for example, there is a cluster of active fractures in the bottom 18m which correlate well with temperature anomalies, as well as two fractures between 30-32m elevation and one at 44m. The anomalies are on the order of ±0.05 °C, with generally smaller anomaly magnitudes in the upgradient section AA'. Anomalies can be positive or negative, depending on the surface temperature history (i.e. cycle time) and connectivity of the local fracture network. Of particular interest are the anomalies at depth, which are found within a few metres of the bottom boundary. Profiles 250 days later (Figure 7) show nearly identical anomalies below about 35m elevation, but show significant differences from 35m to ground surface.

A second simulation was also run assuming a fracture network with a mean aperture of 124µm, which was derived for the Cambridge, Ontario site based on measured packer-transmissivity tests (Burns, 2005). Although the horizontal gradient was maintained at 0.005, surface recharge was reduced by one-half (to 200 mm/yr), to compensate for the lower bulk hydraulic conductivity of the dolostone which would otherwise cause watertable mounding. The 2D temperature distribution of this simulation (not shown) was more homogeneous than the 500µm (mean aperture) case, with near-horizontal temperature contours and essentially no visible temperature anomalies in the raw data.

The temperature (T), temperature difference (ΔT) and velocity profiles in the 124 µm (mean aperture) case, are provided in Figure 8. With lower fracture velocities due to the smaller apertures, there are fewer temperature anomalies and less correlation between the ΔT
and fracture velocity profiles. However, there is a clear fracture-induced anomaly in section BB' at an elevation of 16-18m, and there is significantly more small-scale detail in the max and min $\Delta T$ between the 30m and 45m elevations, respectively. This higher detail is likely a result of less numerical dispersion with the lower fracture velocities. As in the previous case, some variation in T and $\Delta T$ arises from fractures upgradient of the profile location but which do not intersect, and hence are not shown.

![Diagram](image)

Figure 6. Simulated temperature and fracture velocity profiles at (a) Section AA' and (b) Section BB', extracted from the 2D simulation at a cycle time of $t_c=105$ days (see Fig 5a). $Temp$ represents the true simulated temperature, while $\Delta T$ represents the temperature difference between the true profile and the same profile smoothed using a boxcar-filter.

![Diagram](image)

Figure 7. Simulated temperature and fracture velocity profiles at (a) Section AA' and (b) Section BB', extracted from the 2D simulation at a cycle time of $t_c=355$ days (see Fig 5c), i.e. 250 days following the profiles shown in Fig. 6.
Figure 8. Simulated temperature and fracture velocity profiles at (a) Section AA’ and (b) Section BB’, extracted from a 2D simulation (not shown) assuming a network with a mean fracture aperture of 124 µm; cycle time is t_c=105 days (compare with Fig. 6). Temperature change profiles (ΔT) are shown at two scales to highlight response.

Conclusions

As in an unfractured porous medium, heat transport within a fractured porous rock is governed by thermal convection and conduction within the fracture network and porous rock matrix. Thermal conduction between active fractures and the porous matrix can lead to high thermal retardations (up to 500 in the single-fracture case after 10 days) and can significantly attenuate thermal signals.

Nevertheless, the numerical results shown herein clearly show that temperature signals from natural surface variations have the potential to penetrate deep within a fractured rock. Anomalies at depths of 40-50m were clearly correlated with active fractures in the simulations. Although often weak, in these cases on the order of ±0.05 to ±0.002°C, these temperature anomalies are well within the sensitivity range of new, continuous profiling temperature probes being developed for lined boreholes (Pehme et al., 2007).

Assuming a fracture network with a mean fracture aperture of 500 µm, and a hydraulic gradient of 0.005, excellent correlations were obtained between the locations of active fractures and the simulated temperature anomalies. Because of lower fracture velocities, correlations were less evident in a simulation assuming a mean aperture of 124 µm.

The discrete fracture network model SMOKE R has been shown useful for simulating heat transport within fractured rock. Further simulations are underway to evaluate temperature anomalies due to transient recharge events and signatures from artificial thermal sources and thermal tracers. Whether such temperature anomalies can be detected in the presence of background thermal noise remains to be verified.
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John A. Cherry is a Distinguished Professor Emeritus at the University of Waterloo, where he had been a faculty member from 1971-2006. His research focuses on field studies of contaminants in groundwater and the development of monitoring techniques to provide improved insight into contaminant behaviour and fate in granular and fractured media. He co-authored the textbook “Ground Water” with R.A. Freeze (1979) and co-edited and co-authored several chapters in the book “Dense Chlorinated Solvents and Other DNAPLs in Groundwater” (1996). He held the NSERC Industrial Research Chair in Contaminant Hydrogeology at the University of Waterloo (1996-2006) and is currently the Director of the University Consortium for Field-Focused Groundwater Contamination Research, established in 1988.

Beth L. Parker has a Bachelors degree in environmental science/ economics from Allegheny College, a Masters degree in environmental engineering from Duke University and a Ph.D. in hydrogeology from the University of Waterloo. She was a research faculty member in the Earth Sciences Department at the University of Waterloo from 1996 to 2007. She is currently a professor in the School of Engineering at the University of Guelph and holder of the NSERC Industrial Research Chair in Fractured Rock Contaminant Hydrology. Her research involves field studies of transport, fate and remediation of chlorinated solvents in diverse hydrogeologic environments including fractured rock, clayey aquitards and sandy aquifers.