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A dynamic heat transfer coefficient between fractured rock and flowing fluid

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ABSTRACT

Estimation of heat production remains a major challenge for geothermal industry. In continuum mechanics two main approaches need to be separated to model heat transfer between fluid and rock: local thermal equilibrium (LTE) and local thermal non-equilibrium (LTNE). While LTNE does not require the strong assumption of instantaneous local thermal equilibrium, the parameters for explicit heat transfer between rock and fluid are only loosely defined. This work focuses on the heat transfer coefficient between rock walls and flowing fluid. Based on an experimental setup with simple geometry and a steady state scenario, we derive a dynamic heat transfer coefficient dependent on fracture aperture, flow velocity and thermal parameters. We compare our model to experimental data and achieve a good agreement for most temperatures. In comparison to a static heat transfer coefficient, a dynamic coefficient changes the fluid and rock temperature distribution in the fractured system. We then show possible extensions of our dynamic approach with a simulation on reservoir scale. In opposite to existing models and empiric approaches our model intrinsically adjusts to spatial heterogeneity and temporal changes in flow and temperature field. The model is based on well-defined physical parameters which can be easily obtained from standard laboratory tests and dependent on characteristic variables like velocity and rock temperature. Our model can be extended by including more constitutive relationships linking permeability, fracture aperture, fluid pressure and heat transfer.

1. Introduction

Numerical simulations of geothermal systems during stimulation and production phases are common prognostic tools nowadays (e.g. Kolditz et al., 2012; Xu et al., 2006; Kolditz and Clauser, 1998). Complex fluid flow models consider phase change of the fluid or dynamic behavior of permeability and porosity (Rutqvist et al., 2001; Pruess, 2002, 2006). Yet, estimation of heat production remains a major challenge. Some of the issues related to heat transport can be addressed with a continuum approach.

Two models need to be separated in this context: In the equivalent temperature model, the fluid and rock are represented as a single continuum and a common temperature is calculated, assuming that fluid and rock reach local thermal equilibrium (LTE) instantaneously. In the context of geothermal systems and other applications this assumption might not be fully applicable as a temperature gradient between fluid and rock is essential for the system. Also a recent study shows significant differences between LTE and a method neglecting local thermal equilibrium (LTNE) for long-term production (Shaik et al., 2011). Al-Sumaily and Thompson (2013) show that air temperature during flow in a porous cylinder at high flow rates can also not described well by LTE. If local thermal equilibrium is neglected, heat exchange between fluid and rock is calculated using a transfer term, which depends on specific contact area A between rock and fluid, on the temperature difference $\Delta T = T_f - T_r$ between rock and fluid and heat transfer coefficient h. The transferred heat Q is then calculated as (Nield and Bejan, 2013):

$$Q = hA\Delta T \quad (1)$$

The input parameter h is, amongst others, influenced by the geometry of the fluid–rock interface and the fluid motion. So far, different approaches exist to estimate the heat transfer coefficient. One method is the use of a thermal boundary layer, assuming laminar flow along a plate with constant temperature (Chapman, 1989). Values calculated with this method are found to be rather high (Zhao, 2014) and are constant in time and space like all other available approaches. For a porous medium the heat transfer coefficient can be calculated with an empirical correlation using the Nusselt

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number and, assuming spherical particles, the spherical diameter of the grains as described in Dixon (1988). For more complex particle geometries or a fracture system, so far a satisfying expression for h is missing. In most geothermal systems heat transfer and fluid flow primarily take place within fractures. We therefore focus on fracture geometries in this study.

A series of experiments is presented in Zhao and Tso (1993) to estimate the heat transfer coefficient for a steady state case. Fluid flows through a single fracture in a rock specimen with known inflow temperature, while the rock specimen is heated from the outside. Fluid velocity, fracture aperture and temperatures of rock and fluid are varied over 78 different experiments. With the observed fluid outflow temperature, the authors present empirical power-laws concerning the dependence of h on velocity and fracture aperture (Zhao and Tso, 1993). The study of Zhao and Tso (1993) is one of the very few conducted to study the influence of different parameters on the heat transfer coefficient. Recently Zhao (2014) renew the interpretation of these experiments by combining the experimental results with analytical solutions of heat flow in the experimental setup and was able to derive an analytical equation for the heat transfer coefficient for the steady state case. One drawback of the solution presented in Zhao (2014) is the specific adaption to the experiment, which makes it difficult to utilize the approach in a prospective geothermal simulation. This applies especially to the choice of parameters and the steady state assumption. Further h is assumed to be constant in time and space, similar to other previous studies (e.g. Shaik et al., 2011; Dixon, 1988; Staniforth and Cote, 1991). Experimental data and analytical models show dependencies between h and other system variables and parameters like flow velocity, fracture aperture and temperature which are not constant in realistic applications. Also previous studies assume a constant effective value for a whole system, not considering local heterogeneity like it may occur in permeability in a fractured system.

In this work we will examine the solution derived in Zhao (2014) in more detail and derive an equation for a locally defined, dynamic heat transfer coefficient, overcoming the shortfalls mentioned above. This model is suitable for a dynamic, evolving heat transport scenario in a fractured environment and usable in a simulation of a geothermal system. We test our approach with experimental data from Zhao and Tso (1993) and present a study on reservoir scale to prove the scalability of this model.

2. Theory

We consider a single horizontal fracture with aperture $2b$ inside an impermeable specimen with length L = 102 mm and width $2R$ = 51 mm (see Fig. 1). This corresponds to the experimental setup in Zhao and Tso (1993). Water is injected into the fracture on the left side with temperature T_{in} and leaves the fracture on the right with temperature T_{out}. The specimen is heated at top and bottom with

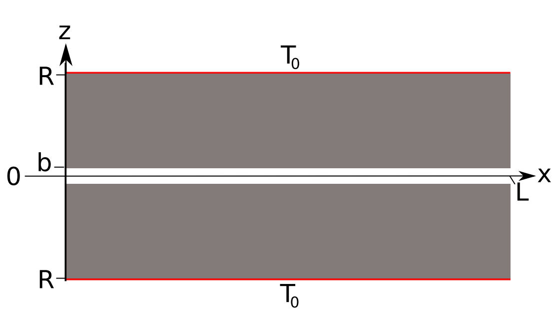

Fig. 1. Sketch of the experimental setup used in Zhao and Tso (1993) and Zhao (2014) and for derivation of the analytical solution.

temperature T_0. In total 78 experiments with different fluid velocities, temperatures and apertures were conducted (see Table 1). Fluid velocities are assumed to be uniform and constant in the fracture.

The governing equation for rock temperature T_r assuming local thermal non-equilibrium is

$$\frac{\partial T_\mathrm{r}}{\partial t} = \frac{K_\mathrm{r}}{(1-\phi)\rho_\mathrm{r} c_\mathrm{r}}\nabla^2 T_\mathrm{r} + \frac{1}{(1-\phi)\rho_\mathrm{r} c_\mathrm{r}} Q \tag{2}$$

where K_α is thermal conductivity, ρ_α is density, ϕ is porosity and c_α heat capacity of phase $\alpha = \{\mathrm{r}, \mathrm{f}\}$ (Nield and Bejan, 2013). The heat equation of the fluid considers an additional advection term with velocity v.

$$\frac{\partial T_\mathrm{f}}{\partial t} = \frac{K_\mathrm{f}}{\phi\rho_\mathrm{f} c_\mathrm{f}}\nabla^2 T_\mathrm{f} - \frac{1}{\phi\rho_\mathrm{f} c_\mathrm{f}} Q - \frac{1}{\phi} v \nabla(T_\mathrm{f}) \tag{3}$$

For the steady state case the rock temperature can be simplified, assuming that heat conduction only takes place perpendicular to the plane (Zhao, 2014).

$$\frac{\partial^2 T_\mathrm{r}}{\partial z^2} = 0 \tag{4}$$

With the boundary condition at the fracture surface $z = b$

$$K_\mathrm{r}\frac{\partial T_\mathrm{r}(x,z)}{\partial z} = -h(T_\mathrm{f}(x,b) - T_\mathrm{r}(x,b)), \tag{5}$$

and the boundary condition at top and bottom $z = R$

$$T_\mathrm{r}(x,R) = T_0, \tag{6}$$

an analytical expression for rock temperature depending on fluid temperature can be derived:

$$T_\mathrm{r}(x,z) = \frac{h}{K_\mathrm{r} + hR}(T_0 - T_\mathrm{f}(x))(z-R) + T_0 \tag{7}$$

As shown in Zhao (2014), conductivity in the water can be neglected and water flow is only one-dimensional in the experimental setup (cf. Fig. 1). Using Eq. (7) and the boundary conditions

$$T_\mathrm{f}(x=0) = T_{in} \tag{8}$$

$$T_\mathrm{f}(x=\infty) = T_0 \tag{9}$$

an analytical solution for fluid temperature in a steady state is derived

$$T_\mathrm{f}(x) = T_0 + (T_{in} - T_0)\exp\left(-x\frac{hAK_\mathrm{r}}{v\rho_\mathrm{f} c_\mathrm{f}(K_\mathrm{r} + hR)}\right) \tag{10}$$

For a known fluid temperature at position x the heat transfer coefficient can be calculated

$$h = -\frac{v\rho_\mathrm{f} c_\mathrm{f} K_\mathrm{r}\ln\left(\frac{T_\mathrm{f}(x)-T_0}{T_{in}-T_0}\right)}{xAK_\mathrm{r} + v\rho_\mathrm{f} c_\mathrm{f} R\ln\left(\frac{T_\mathrm{f}(x)-T_0}{T_{in}-T_0}\right)} \tag{11}$$

In the study of Zhao (2014) Eq. (11) was used to calculate a constant, homogeneous heat transfer coefficient for the experimental setup in a steady state with $T_\mathrm{f}(x = L) = T_\mathrm{out}$. Several improvements need to be applied to expand this idea to calculate such an effective heat transfer coefficient for geothermal applications. First of all, this solution contains parameters, which are problem specific to the experimental setup. In the experiment, parameter R describes the shortest distance from the fracture to the heating plate, or more general, the distance to the defined temperature T_0. In the experimental setup and the analytical solution, this is the only known rock temperature. In the context of a geothermal system, R could be interpreted as some characteristic length scale of the system in terms of an influence domain, giving the shortest distance to a point where rock temperature remains constant through fluid