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Thermal coupling may control mechanical stability of geothermal reservoirs during cold water injection

Silvia De Simone [a,b,*], Victor Vilarrasa [a,b], Jesús Carrera [a], Andrés Alcolea [c], Peter Meier [c]

[a] GHS, Institute of Environmental Assessment and Water Research (IDAEA), Consejo Superior de Investigaciones Científicas (CSIC), Jordi Girona 18-26, 08034 Barcelona, Spain
[b] GHS, Department of Geotechnical Engineering and Geosciences, Technical University of Catalonia (UPC-BarcelonaTech), Jordi Girona 1-3, 08034 Barcelona, Spain
[c] Geo-Energie Suisse AG, Reitergasse 11, CH-8004 Zürich, Switzerland

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ABSTRACT

Hydraulic stimulation and geothermal reservoir operation may compromise the rock mechanical stability and trigger microseismic events. The mechanisms leading to this induced seismicity are still not completely understood. It is clear that injection causes an overpressure that reduces the effective stress, bringing the system closer to failure conditions. However, rock instability may not result only from hydraulic effects, but also from thermal effects. In fact, hydro-mechanical (i.e., isothermal) models often fail to reproduce field observations because the injection of cold water into a hot reservoir induces thermal stresses due to rock contraction. Thus, rock instability is likely to result from the superposition of hydraulic and thermal effects. Here, we perform coupled thermo-hydro-mechanical and hydro-mechanical simulations to investigate the effects of cold water injection in a fracture zone-intact rock system. Results show that thermal effects induce a significant perturbation on the stress in the intact rock affected by the temperature drop. This perturbation is likely to trigger induced seismicity in the surroundings of critically oriented fractures near the injection well. Hydro-mechanical simulations show that the behavior depends on the orientation of the faults and on the initial stress tensor. In the direction of the fractures, where the strains are more constrained, total stress increases with increasing pressure; thus, deviatoric stress increases or decreases depending on the initial stress state. The comparison between hydraulic and thermal effects shows that, when the largest confining stress acts perpendicular to the fractures, thermoelastic effects dominate and could trigger induced seismicity.

1. Introduction

Geothermal energy production from deep hot rocks requires a high permeability heat exchanger for economic efficiency. The typical procedure entails intercepting natural pre-existing discontinuities, such as faults and joints, and enhancing their permeability by means of stimulation. Hydraulic stimulation is the most widely used method. It involves the massive injection of a large volume of water (several thousand cubic meters) at high flow rates to increase the downhole pore pressure, which tends to induce shearing along the fracture planes (Pearson, 1981). In this way permeability is enhanced due to dilatancy, especially in the direction perpendicular to shear (Barton et al., 1985; Yeo et al., 1998; Mallikamas and Rajaram, 2005).

Microseismic events occur during hydraulic stimulation. Induced seismicity is typically weak ($M < 2$; Evans et al., 2012) and certifies the effectiveness of the operation. However, these events are sometimes of sufficient magnitude to be felt by the local population. For example, seismic events with magnitude greater than 3 occurred at the Basel Deep Heat Mining Project in Switzerland (Häring et al., 2008) and at the Hot Dry Rock Project of Soultz-sous-Forêts in France (Cornet et al., 1997; Baria et al., 2005). This causes a negative impact on the local population and may compromise the continuation of the project. Hence, understanding the mechanisms triggering these induced micro-earthquakes is important to properly design and manage geothermal stimulation and operation so as to prevent them.

Induced seismicity occurs when failure conditions are reached, either at an existing fracture or at a newly created one. It is widely believed that the main cause of failure during hydraulic stimulation is overpressure (Shapiro et al., 1999, 2003; Parotidis et al., 2004). Indeed, overpressure produces a reduction of effective stresses that can cause the fracture to yield. Rutqvist and Stephansson (2003) provide an accurate review of the hydro-mechanical coupling in fractured rock and point out its relevance in the geothermal field. However, pore pressure cannot be considered the only cause of induced seismicity. Microseismic events at Soultz-sous-Forêts (Baisch et al., 2010; Evans et al., 2005) and Basel (Häring et al.,

* Corresponding author at: GHS, Institute of Environmental Assessment and Water Research (IDAEA), Consejo Superior de Investigaciones Científicas (CSIC), Jordi Girona 18-26, 08034 Barcelona, Spain. Tel.: +34 665737253.
E-mail address: silviadesi@gmail.com (S. De Simone).

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2008; Ripperger et al., 2009) were still occurring once injection stopped and often the largest microseisms occurred after the end of injection, like in Basel. These post injection events cannot be explained by pressure diffusion alone, because its magnitude decreases quickly with time (Delleur, 1999).

Interestingly, injected water was cold both at Basel and Soultz. The temperature contrast between the hot reservoir and the injected water (at atmospheric conditions at surface) was large. This produces a significant contraction of both the fracture filling and the surrounding rock, leading to an additional reduction in effective stresses, which has to be taken into account (Majer et al., 2007). This effect is confirmed by the measurements taken at the Geysers geothermal steam field (Santa Rosa, California), where the observed seismicity is not directly related to overpressure (National Research Council, 2012). There, the large temperature difference between the injected fluid and the deep rock produced a significant cooling of the geothermal reservoir (Mossop and Segall, 1997), which caused thermal contractions of the rock, affecting the in situ stress state. In short, thermal effects should be considered to better understand the processes involved in geothermal reservoir stimulation (Segall and Fitzgerald, 1998). To achieve this, coupled thermo-hydro-mechanical analyses are necessary.

Thermoelastic effects in geothermal systems have been studied by some authors (e.g. Ghassemi, 2012). They performed thermo-hydro-mechanical models of cold water injection into a planar fracture (Kohl et al., 1995; Ghassemi et al., 2007, 2008; Ghassemi and Zhou, 2011) or in a fracture network (Kolditz and Clauser, 1998; Bruel, 2002; McDermott et al., 2006). Nevertheless, most of them solely point out the perturbations generated within the fracture or at the fracture's surface, but not the effects on the surrounding rock mass.

We conjecture that the thermal effects developing in the cooled part of the rock matrix may play an important role in triggering induced seismicity. In fact, the intact rock has a greater stiffness than the fracture, so that thermal stress changes may indeed be large, which could explain how seismic events are triggered. To investigate this, we simulate the hydraulic stimulation of an idealized fracture zone embedded in an intact rock matrix. Hydraulic and thermal effects are studied by means of fully coupled thermo-hydro-mechanical (THM) simulations of cold water injection into the hot fracture zone/matrix system. The results of the THM simulation are compared to those of a hydro-mechanical (HM) simulation (i.e. injection of water in isothermal equilibrium with the hot rock) in order to estimate the impact of the thermal effects.

2. Methods

2.1. Conceptual model

To investigate the effect of the cooling front caused by cold water injection on thermoelastic strain, we perform coupled HM and THM numerical simulations of water injection into a rock mass containing a zone of discontinuities. An idealized geometry consisting of a planar fracture zone (corresponding to joints or faults) of 1 m thickness embedded into an intact rock mass is considered. The fracture zone is treated like a continuous porous medium. This assumption is made considering that faults often consist of a fault core with a thickness of few centimeters embedded into a highly damaged zone of some tens of centimeters (Gudmundsson, 2004; Cappa and Rutqvist, 2011). The whole fracture zone is surrounded by the host rock, which is generally less permeable and stiffer than the fracture zone. In fractured crystalline rocks the intrinsic permeability of the intact rock matrix may be some 5 orders of magnitude smaller than that of the fault zone (Rutqvist and Stephansson, 2003). This difference in hydraulic properties converts the fault zone into a preferential flow path.

The numerical simulations calculate deformations and changes in the stress field due to cold water injection. Linear elasticity is assumed for the whole model. In order to evaluate the potential induced seismicity, we perform a slip tendency analysis (Byerlee, 1978; Morris et al., 1996; Streit and Hillis, 2004). We consider the Mohr–Coulomb failure criterion (Jaeger et al., 2007)

$$\tau_r = c + \mu \cdot \sigma'_n \tag{1}$$

where τ_r is the critical shear stress, c is cohesion, σ'_n is effective normal stress and μ is friction coefficient, which is often expressed in term of the angle of friction ϕ ($\mu = \tan\phi$).

For cohesion-less materials ($c = 0$), sliding occurs when the shear stress τ equals the critical shear stress τ_r, i.e. when the ratio of the shear stress to effective normal stress equals the friction coefficient μ

$$\frac{\tau}{\sigma'_n} = \tan\phi \tag{2}$$

We use this equation to estimate the mobilized friction angle ϕ_{mob} on critically oriented planes (i.e. the one for which ϕ_{mob} is maximum). This value quantifies the shear slip tendency along a plane, because it represents how close is the stress state to the failure envelope.

2.2. Mathematical model

According to linear theory of poro-thermoelasticity (McTigue, 1986), stresses are a function of strain, fluid pressure and temperature

$$\Delta\boldsymbol{\sigma} = K\varepsilon_v\mathbf{I} + 2G\left(\boldsymbol{\varepsilon} - \frac{\varepsilon_v}{3}\mathbf{I} + \frac{1}{2G}\Delta p_f\mathbf{I} - \frac{3K}{2G}\alpha_T\Delta T\mathbf{I}\right) \tag{3}$$

where $\boldsymbol{\sigma}$ is the total stress tensor, ε_v is volumetric strain, $\mathbf{I}$ is the identity matrix, $\boldsymbol{\varepsilon}$ is the strain tensor, $K = E/(3(1-2v))$ is the bulk modulus, $G = E/(2(1+v))$ is the shear modulus, E is the Young's modulus, v is Poisson ratio, p_f is the fluid pressure, α_T is the linear thermal expansion coefficient and T is temperature. Biot coefficient has been assumed to be 1 because the rock compressibility is negligible compared to that of the grains. Moreover, this value is the least favorable, because it leads to the strongest hydromechanical coupling (see also Zimmerman, 2000). Notice that a temperature drop implies an isotropic drop in stresses equal to $3K\alpha_T\Delta T$, which can be very large for stiff rocks.

To solve the mechanical problem, the momentum balance has to be satisfied. If the inertial terms are neglected, it reduces to the equilibrium of stresses

$$\nabla \cdot \boldsymbol{\sigma} + \mathbf{b} = \mathbf{0} \tag{4}$$

where $\mathbf{b}$ is the vector of body forces.

Eq. (3) is coupled with the flow equation through fluid pressure. Assuming that there is no external loading and neglecting solid phase compressibility, fluid mass conservation of the fluid can be written as

$$\frac{\phi}{K_f}\frac{\partial p_f}{\partial t} + \nabla \cdot \frac{d\mathbf{u}}{dt} + \frac{1}{\rho}\nabla \cdot (\rho\mathbf{q}) = f_w \tag{5}$$

where ϕ is porosity, $1/K_f$ is water compressibility, t is time, $\mathbf{u}$ is the solid displacement vector, $\mathbf{q}$ is the water flux and f_w is an external supply of water. Notice that the second term represents the rate of change in volumetric strain (i.e. porosity). The water flux is given by Darcy's Law

$$\mathbf{q} = -\frac{k}{\mu(p_f, T)}(\nabla p_f + \rho(p_f, T) \cdot g \cdot \nabla z) \tag{6}$$

where k is the intrinsic permeability, g is gravity, z is the vertical coordinate, and μ and ρ are respectively the fluid viscosity and