Geothermics 51 (2014) 130–141

Contents lists available at ScienceDirect

Geothermics

journal homepage: www.elsevier.com/locate/geothermics

Reactive transport modeling of the Dixie Valley geothermal area: Insights on flow and geothermometry

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ARTICLE INFO

Article history:
Received 6 September 2013
Accepted 16 December 2013
Available online 15 January 2014

Keywords:
Reactive transport modeling
Solute geothermometry
Dixie Valley
Fracture flow
Geothermal springs

ABSTRACT

A 2D reactive transport model of the Dixie Valley geothermal area in Nevada, USA was developed to assess fluid flow pathways and fluid rock interaction processes. The model includes two major normal faults and the incorporation of a dual continuum domain to simulate the presence of a small-scale thermal spring being fed by a highly permeable but narrow fracture zone. Simulations were performed incorporating fluid flow, heat conduction and advection, and kinetic mineral-water reactions. Various solute geothermometry methods were applied to simulated spring compositions, to compare estimated reservoir temperatures with "true" modeled reservoir temperatures, for a fluid ascending the simulated fracture and cooling on its way to the surface. Under the modeled conditions (cooling but no mixing or boiling), the classical Na–K(–Ca) geothermometers performed best because these are least affected by mineral precipitation upon cooling. Geothermometry based on computed mineral saturation indices and the quartz geothermometer were more sensitive to re-equilibration upon cooling, but showed good results for fluid velocities above ca. 0.1 m/d and a reactive fracture surface area 1–2 orders of magnitude lower than the corresponding geometric surface area. This suggests that such upflow rates and relatively low reactive fracture surface areas are likely present in many geothermal fields. The simulations also suggest that the presence of small-scale fracture systems having an elevated permeability of 10^{-12} to 10^{-10} m^2 can significantly alter the shallow fluid flow regime of geothermal systems. For the Dixie Valley case, the model implies that such elevated permeabilities lead to a shallow (less than 1 km) convection cell where superficial water infiltrates along the range front normal fault and connects the small-scale geothermal spring through basin filling sediments. Furthermore, we conclude that a fracture permeability on the order of 10^{-12} m^2 may lead to near surface temperature >100 °C whereas a permeability of 10^{-10} m^2 is not realistic because this permeability led to extreme upflow velocities and to a short-circuit of the regional fault zone.

1. Introduction

Although simulations of geothermal systems have in some cases incorporated reactive transport (Xu and Pruess, 2001; Dobson et al., 2003, 2004) most large-scale (2–3D) models for geothermal areas have only taken into account fluid flow and heat transfer (e.g., Clearwater et al., 2012; McKenna and Blackwell, 2004; Moulding and Brikowski, 2012). Fully coupled reactive transport models of field scale geothermal systems are rarely found in the literature. Exceptions are simulations of enhanced geothermal systems (EGS) (Bachler and Kohl, 2005; Sonnenthal et al., 2012; Taron and Elsworth, 2009), formation of scale within geothermal wells (Alt-Epping et al., 2013; Xu et al., 2004) or the simulation of shallow hydrothermal systems (Jones and Xiao, 2006; Xu and Pruess, 2001).

In this study, a 2D reactive transport model of the Dixie Valley geothermal area (Nevada, USA) was developed to assess fluid flow pathways and fluid rock interaction processes. The Dixie Valley geothermal field, located in the Basin and Range province of the western US, was chosen as an example study because it has been used for power production (ca. 63 MW) over the last two decades and has been extensively characterized (Blackwell et al., 2007, and references therein). Our reactive transport model specifically benefits from the availability of an extensive geochemical and isotopic dataset (Goff et al., 2002). Field scale features include geothermal springs with temperatures up to 84 °C (Goff et al., 2002), subsurface temperatures in excess of 280 °C at 3 km depth, the absence of known magmatic heat sources and an elevated basal

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http://dx.doi.org/10.1016/j.geothermics.2013.12.003

heat flux on the order of 90 mW/m^2 which is typical for the Basin and Range province (McKenna and Blackwell, 2004). Recent investigations also showed that near surface groundwater temperatures can be greater than 100 °C for isolated locations (Iovenitti et al., 2012). The general understanding of Basin and Range geothermal systems is that meteoric water infiltrates via the range top or valley floor, heats up during deep circulation and ascends along the most permeable pathways such as range-bounding normal faults (McKenna and Blackwell, 2004).

To what depth the fluid circulation extends, however, is still under debate. Helium isotopic studies revealed that ca. 7.5% of the He in the Dixie Valley system is derived from mantle sources, requiring fluid input from below the brittle–ductile transition (Kennedy and van Soest, 2006). McKenna and Blackwell (2004) postulated a large scale fluid convection cell where infiltrating meteoric water reaches a depth up to 8 km before it finally ascends to the surface. Moulding and Brikowski (2012), on the contrary, argued that such deep fluid infiltration seems unrealistic considering that the lithostatic stress at this depth reduces the permeability needed to establish significant advective fluid flow (5×10^{-17} m^2, McKenna and Blackwell, 2004). Reduced permeability with increasing depth is especially enhanced below the brittle–ductile transition such as shown by Weis et al. (2012). For a temperature gradient of ca. 115 °C/km, Weis et al. (2012) illustrate that this transitions occurs at a depth of ca. 3–5 km. Blackwell et al. (2007) showed that the average temperature gradient for the Dixie Valley area is 63 °C/km. For isolated locations, however, the temperature gradient reaches values that are much greater than 100 °C/km (Blackwell et al., 2007). These high values suggest that the brittle–ductile transition and accompanying permeability reduction might occur at relatively shallow depth at Dixie Valley. To account for the potentially low rock permeability, Moulding and Brikowski (2012) postulate that the fluid circulation is much shallower when compared to the model of McKenna and Blackwell (2004) by presenting a full 3D model that accounts for horizontal fluid flow within a large-scale fracture plane.

In contrast to the flow simulation studies performed by McKenna and Blackwell (2004) and Moulding and Brikowski (2012), the main objective of our reactive transport modeling study is to evaluate the controlling attributes of hydrothermal convection (e.g. fluid velocity, fracture surface areas, etc.) and their impact on surface fluid chemistry and the use of solute chemical geothermometry.

Geothermometry based on mineral saturation indices has been used as an exploration tool for geothermal systems (Reed and Spycher, 1984; Spycher et al., 2011, 2014). This method, referred to here as multicomponent geothermometry, involves computing multicomponent homogenous equilibria to yield temperature-dependent saturation indices of selected potential reservoir minerals, using full water chemical analyses. The clustering of saturation indices near zero (the equilibrium point) at any temperature, for a group of reservoir minerals, provides an indication of probable reservoir temperature. This method is different than the classical and empirical SiO_2 and Na–K–Ca geothermometers (Fournier and Rowe, 1966; Fournier and Truesdell, 1973) because it relies on more than a few selected chemical components. Both multicomponent and classical geothermometry methods are based on the assumption that the chemical compositions of waters sampled at the surface reflect equilibrium with reservoir minerals at depth. An advantage of multicomponent geothermometry is that it can yield reservoir temperatures in systems that are approaching equilibrium, although not necessarily at equilibrium, and can also be used to deconvolute effects from degassing and mixing (Spycher et al., 2014).

The potential for re-equilibration of geothermal fluids as they ascend from the deep reservoir to the surface is a shortcoming of chemical geothermometry, resulting in the "deep" chemical and temperature signature of the fluid being lost, and thus causing erroneous temperature estimates (Ferguson et al., 2009). To address the re-equilibration issue, we evaluated the necessary upflow velocity along a spring-feeding fracture and the minimum corresponding reactive fracture surface area needed to avoid re-equilibration upon conductive cooling. Varying upflow velocity and fracture reactivity in our model allowed us to define flow and reactive constraint conditions under which geothermometry is expected to be most successful. Moreover, our model simulations provide insight into fluid circulation depths and upflow velocities for typical Basin and Range geothermal systems.

2. Model setup and calibration

2.1. Numerical simulator

All simulations were performed using the newly parallelized version of TOUGHREACT (Sonnenthal et al., http://esd.lbl.gov/research/projects/tough/software/toughreact.html) based on TOUGHREACT V2 (Xu et al., 2011), allowing a computationally efficient simulation of fully coupled reactive transport in variably saturated geologic media. TOUGHREACT is based on the TOUGH2 simulator (Pruess et al., 1999) that simulates fluid flow and heat transfer processes. For this study we used the TOUGH2 equation of state module EOS1 considering fully saturated, non-isothermal water flow occurring as a single phase only. By simulating one single fluid phase, the model does not consider boiling. This constraint forms a model simplification because there are a few fumaroles, and thus boiling occurring at Dixie Valley (McKenna and Blackwell, 2004). Neglecting boiling is, however, justified because our main objective was to assess the effects of conductive cooling during varying upflow velocities as well as fracture reactivity on the use of chemical geothermometry. Moreover, Spycher et al. (2014) and Peiffer et al. (2014) recently discussed in detail how boiling can be taken into account for temperetures estimates using multicomponent geothermometry.

TOUGHREACT simulates water and heat flow using the governing mass balance equations (here shown for single-phase water only):

$$\frac{\partial M_{W,H}}{\partial t} = -\nabla F_{W,H} + q_{W,H} \tag{1}$$

where $M_{W,H}$ is the accumulation term for water M_W (kg/m^3) or heat M_H (J/m^3), $q_{W,H}$ are water or heat sinks (−) or sources (+) and $F_{W,H}$ refers to the water flux F_W (kg m^{-2} s^{-1}) or heat flux F_H (J m^{-2} s^{-1}). For fully saturated, single phase flow problems F_W is equal to the Darcy flux u (m/s)

$$u = -\frac{k}{\mu}(\nabla P - \rho g) \tag{2}$$

where k refers to the permeability (m^2), μ is the water viscosity (kg/m/s), P (Pa) refers to the water pressure and ρ and g are the water density (kg/m^3) and the gravitational acceleration (m/s^2), respectively. Heat flux F_H (J s^{-1} m^{-2}) is defined as:

$$F_H = C_M \cdot T \cdot \rho_M \cdot u - \lambda \cdot \nabla T \tag{3}$$

where λ refers to the wet thermal conductivity of the solid/rock (J s^{-1} m^{-1} K^{-1} = W m^{-1} K^{-1}), T (K) is the temperature of the porous media (rock + water), C_M (J kg^{-1} K^{-1}) and ρ_M (kg/m^3) refer to the specific heat and density of the porous media, and ∇T (K/m) refers to the temperature gradient between solids of adjacent grid blocks.