Supplementary materials for:

Deep permeable fault–controlled helium transport and limited mantle flux in two extensional geothermal systems in the Great Basin, United States

1Amlan Banerjee, 1Mark Person, 2Albert Hofstra, 2Donald Sweetkind, 3Denis Cohen, 4Andrew Sabin, 5Jeff Unruh, 6George Zyvoloski, 6Carl W Gable, 7Laura Crossey and 7Karl Karlstrom

1Department of Earth and Environmental Sciences, New Mexico Tech, Socorro, NM 87801
2United States Geological Survey, Denver Federal Center, Denver, CO 80225
3Department of Geological & Atmospheric Sciences, Iowa State University, Ames, Iowa 50011
4Navy Geothermal Program Office, China Lake, CA 93555
5William Letts & Associates, 1777 Botelho Dr., Suite 26, Walnut Creek, CA 94596
6Computational Earth Science Group, Los Alamos National Laboratory, Los Alamos, NM 87545
7Department of Earth and Planetary Sciences, University of New Mexico, Albuquerque, NM 87131

Table DR1. Thermal Rock Properties

<table>
<thead>
<tr>
<th>Variable</th>
<th>Symbol</th>
<th>Value/Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Porosity</td>
<td>$\phi$</td>
<td>0.05</td>
</tr>
<tr>
<td>Thermal Conductivity of Fluid</td>
<td>$\lambda_\text{f}$</td>
<td>0.58 W-m°C</td>
</tr>
<tr>
<td>Thermal Conductivity of Solids</td>
<td>$\lambda_\text{s}$</td>
<td>2.5 W-m°C</td>
</tr>
<tr>
<td>Heat Capacity of Fluid Phase</td>
<td>$c_\text{f}$</td>
<td>4180 J/kg</td>
</tr>
<tr>
<td>Heat Capacity of Solid Phases</td>
<td>$c_\text{s}$</td>
<td>800 J/kg</td>
</tr>
<tr>
<td>Longitudinal Dispersivity</td>
<td>$\alpha_\text{L}$</td>
<td>1.0 m</td>
</tr>
<tr>
<td>Transverse Dispersivity</td>
<td>$\alpha_\text{T}$</td>
<td>0.1 m</td>
</tr>
<tr>
<td>Solute Diffusivity</td>
<td>$D_\text{d}$</td>
<td>$10^{-10}$ m$^2$/s</td>
</tr>
</tbody>
</table>

Table DR2. Hydrologic Rock Properties (Coso geothermal system)

<table>
<thead>
<tr>
<th>log ($k_x$)</th>
<th>log ($k_z$)</th>
<th>$\bar{A}$</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>-16</td>
<td>-16</td>
<td>0.0001</td>
<td>Seismogenic basement</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Low permeability faults</td>
</tr>
<tr>
<td>-19</td>
<td>-19</td>
<td>0.0001</td>
<td>Alteration zone (steam cap)</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Volcanic rocks</td>
</tr>
<tr>
<td>-16</td>
<td>-15</td>
<td>0.0001</td>
<td>Coso Wash fault</td>
</tr>
<tr>
<td>-18</td>
<td>-18</td>
<td>0.000001</td>
<td>Quaternary basin fill</td>
</tr>
<tr>
<td>-16</td>
<td>-16</td>
<td>0.000001</td>
<td>Mesozoic rocks</td>
</tr>
</tbody>
</table>
Table DR3. Hydrologic Rock Properties (Beowawe geothermal system)

<table>
<thead>
<tr>
<th>log (k_x)</th>
<th>log (k_z)</th>
<th>A</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>-13</td>
<td>-12</td>
<td>0.000001</td>
<td>High-angle (Malpais) fault</td>
</tr>
<tr>
<td>-13</td>
<td>-12</td>
<td>0.000001</td>
<td>High-angle (Muleshoe) fault</td>
</tr>
<tr>
<td>-14</td>
<td>-14</td>
<td>0.000001</td>
<td>High-angle faults</td>
</tr>
<tr>
<td>-16</td>
<td>-16</td>
<td>0.000001</td>
<td>Basin fill</td>
</tr>
<tr>
<td>-16</td>
<td>-16</td>
<td>0.000001</td>
<td>Basin fill</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Basin fill</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Volcanic rocks</td>
</tr>
<tr>
<td>-10</td>
<td>-11</td>
<td>0.000001</td>
<td>Carbonate rocks</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Carbonate rocks</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Proterozoic siliciclastic rocks</td>
</tr>
<tr>
<td>-14</td>
<td>-14</td>
<td>0.000001</td>
<td>Proterozoic siliciclastic rocks</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Proterozoic siliciclastic rocks</td>
</tr>
<tr>
<td>-17</td>
<td>-17</td>
<td>0.000001</td>
<td>Proterozoic bed rocks</td>
</tr>
</tbody>
</table>

k_x, k_z (horizontal and vertical fault permeability) – log(m^2), A (surface area) – m^2/mole

Permeability: The permeability values (Tables DR2, DR3) were largely based on model calibration based on observed thermal and isotopic concentration, but are consistent with values reported in the Basin and Range for fractures and sedimentary rocks. Permeability of subvertical faults near Yucca Mountain in the Basin and Range province is estimated to be \(\sim 10^{-10} \text{ m}^2\) (Bredheoef, 1997), while Morin et al., (1998) reported Dixie valley range-bounding fault permeability to range between \(10^{-13.5} - 10^{-11.3} \text{ m}^2\) (assuming a fault width of 20 m). High-permeability conditions along the Basin and Range are due to the high stress state and strain rates of the area (Hickman et al., 1998; Barton et al., 1998; Blewitt et al., 2002). Faulder et al., (1997) estimated Malpais fault (Beowawe) permeability to vary between \(10^{-10.8} - 10^{-11.6} \text{ m}^2\) (assuming a fault width of 20 m). Garg et al. (2007) reported a fault permeability of 200 mili-darcy for the Malpais fault. Davatzes and Hickman (2005) using simple one-dimensional heat flow analytic solution from well data (well 58A-10) reported that the vertical fault permeability in Coso geothermal system ranges from \((10^{-13.5} \text{ to } 10^{-17} \text{ m}^2)\). We run several simulations (not shown here) assigning vertical fault permeability of \(10^{-13} \text{ to } 10^{-16} \text{ m}^2\) for our Coso model. Assigning vertical fault permeability for Coso as \(10^{-15} \text{ m}^2\) keeps the temperature within the wellhead observation by Fournier and Thompson (1980).

Initial and boundary conditions: We specified the heads along the top boundary to reflect water table topography. We assumed no flux conditions along the bottom and sides of the solution domain. A specified value stream function \((\Psi = 0 \text{ m}^2/\text{yr})\) was assigned along the base and sides of the solution domain. A specified flux \((d \Psi/dz)\) proportional to the water table gradient was assigned along the top of the solution domain for the stream function equation. We assumed a conductive geothermal gradient \((30 \degree \text{C/km})\) for initial temperature distribution. Uniform rock isotope values were assigned to stratigraphic units depending on the mineral assemblages and abundances. Temperature dependent equilibrium oxygen fractionation factors were used to assign initial fluid \(\delta^{18} \text{O}\) isotope values. Initial \(\delta \text{D}\) value was fixed at -100 and -140/o/oo for the Coso and
Beowawe respectively. Helium concentrations (³He, ⁴He) within and at the base of the crust is uncertain making the selection of these initial conditions necessarily somewhat arbitrary. For the radiogenic helium (⁴He) we assigned the initial vertical increase in helium composition using Fick’s law [-J/D = (∆C/∆z), where J is the basal ⁴He flux (3.0 × 10⁻¹⁸ mol cm⁻² s⁻¹), D is the diffusivity (10⁻⁶ cm² s⁻¹) and (∆C/∆z) is the vertical concentration gradient]. We included a radiogenic source term in our model for production of ⁴He in the crust (in situ production rate = 5.8 × 10⁻²⁴ mol cm⁻³ s⁻¹; Bethke et al., (1999), Pereira and Adams (1982)). For assigning the initial vertical ³He composition we assumed a basal flux of 1.66 × 10⁻²⁰ mol cm⁻² s⁻¹ and diffusivity of 10⁻⁶ cm² s⁻¹. We specified the ⁴He and ³He such that the basal R/RA value for Coso was fixed at 12. For Beowawe the ³He values were adjusted such that the basal R/RA value varied between 2 and 8 (absence and presence of a magmatic source). The basal helium R/RA values were chosen such that it falls within the range of reported upper mantle helium R/RA ratios (Anderson, 2000a, b, 2002; Meibom et al., 2005; Kurz et al., 1982). Due to the lack of site specific helium flux data and diffusivity of helium through saturated porous media/fault zone, the diffusive ³He and ⁴He fluxes (used to assign initial ³He and ⁴He values in the solution domain) were constrained by the observed R/RA data (Welhan et al., 1988 and Kennedy and van Soest, 2007) from these two sites.

We imposed a specified temperature of 10 °C at the water table. A basal heat flux was imposed at the base of the solution domain consistent with the local heat flow anomaly. Boundary value at the water table for the fluid δ¹⁸O was fixed at -16 ‰. Fluid δD values at the water table were fixed at -100 ‰ and -130 ‰ respectively for Coso and Beowawe. We also assigned the water table ⁴He and ³He values as 2.0 × 10⁻¹² mol cm⁻³ and 2.8 × 10⁻¹⁸ mol cm⁻³ respectively making R/RA = 1. No flux conditions were imposed along the base and sides of the solution domains for oxygen, hydrogen and helium (³He, ⁴He). An advective boundary condition (dT/dz=0, dC¹⁸O/dz=0, dC¹⁰D/dz=0, dC³He/dz=0 and dC⁴He /dz=0) was imposed at the intersection of water table and the discharge faults (the Coso Wash fault and Malpais fault). For the Coso simulation with basal mantle fluid flux we used a flux of 3.32 × 10⁻⁸ kg m⁻² s⁻¹ that falls within the range of typical regional metamorphic fluid flux (10⁻⁹ to 10⁻⁶ kg m⁻² s⁻¹ Hanson, 1992).

The assigned initial and boundary conditions were constrained by measured values and model calibration (i.e., matching measured temperatures, fluid and rock isotopic compositions and helium concentrations). Reactive surface areas were estimated based on fracture density and on observed fluid and rock isotopic alteration measurements. We did not take into account the surface roughness on fracture walls. Reactive surface area of 0.0001 m²/mole corresponds to a relatively close fracture spacing of 50 cm while a coarser fracture spacing of 20.5 m will indicate a reactive surface area of 0.000001 m²/mole (Table DR2, DR3).

**Mathematical equations and simulation methods:** Our model represents both topography-and density-driven flows. Thermodynamic equations of state required to compute the density and viscosity of ground water at elevated temperature, pressure, and salinity conditions are obtained from Haar et al. (1984). We represent stream function equation using the approach outlined in Senger and Fogg (1990). We represent advective-dispersive isotope transport and fluid-rock isotope exchange using the approach outlined in Bowman et al. (1994). For simplification we
assumed that the lithologic units are composed of seven different minerals (derived from Holk and Taylor, 2000; Mulch, 2004). Temperature dependent equilibrium isotope fractionation factors, pre-exponential factor (Ao) and the activation energy (Ea) for each mineral phase were taken from Cole and Ohmoto (1986). We represented transport of $^3$He and $^4$He in porous media using an advective-dispersive formulation. All the governing equations used in this study in presented in Table 1S. For all the transport equations, the resulting set of algebraic equations was solved using Gaussian elimination. We solved variable-density groundwater flow, heat, isotope and solute transport equations (Person et al., 2007) sequentially using the finite element method. To eliminate numerical oscillations associated with convection dominated problems (Pinder and Grey, 1977) the heat and solute transport equations were solved using the modified method of characteristics algorithm (Zheng and Bennett, 1995). A time step size of 10 and 1 year were used for Coso and Beowawe simulation respectively. Each transport equation was solved on a different processor using the open MPI library. This enables the code to run five times faster than a conventional serial code. In order to satisfy the grid Courant number, each scenario was run for tens of thousands of time steps. The simulation time scales were chosen to match field observations. All Coso simulation results presented here are after 200,000 years. In contrast, the Beowawe model was run for only 20,000 years because shallow heat flow data presented by Olmstead and Rush (1987) suggests that this is a young hydrothermal system far from equilibrium. For example, large portions (up to 2/3 the surface area) of Whirlwind Valley (Fig. 1) should be characterized by low heat flow conditions ($< 80 \text{ mW/m}^2$) for long-lived ($> \sim 100,000$ years) hydrothermal systems (Smith and Chapman, 1983) which is not observed. The Beowawe model used a uniform basal heat flow of 80 mW/m$^2$. Basal specified heat flow for the Coso model varied over a range of 60 to 450 mW/m$^2$ between the thickest and thinnest portions of the seismogenic crust.

Table DR4. Governing transport equations:

Variable Density Groundwater Flow:

$$\frac{\partial}{\partial x} \left[ \frac{\rho_0 g}{\mu_f} \left( k_{xx} \frac{\partial h}{\partial x} + k_{xz} \frac{\partial h}{\partial z} \right) \right] + \frac{\partial}{\partial z} \left[ \frac{\rho_0 g}{\mu_f} \left( k_{zx} \frac{\partial h}{\partial x} + k_{zz} \frac{\partial h}{\partial z} \right) \right] = -\rho_r k_{zz} \frac{\rho_0 g}{\mu_f} \tag{S1}$$

Stream Functions:

$$\frac{\partial}{\partial x} \left[ \mu_f \left( \frac{k_{xx}}{k} \frac{\partial \psi}{\partial x} + \frac{k_{xz}}{k} \frac{\partial \psi}{\partial z} \right) \right] + \frac{\partial}{\partial z} \left[ \mu_f \left( \frac{k_{zx}}{k} \frac{\partial \psi}{\partial x} + \frac{k_{zz}}{k} \frac{\partial \psi}{\partial z} \right) \right] = -g \frac{\partial \rho_r}{\partial x} \tag{S2}$$
Heat Transport

\[
\frac{\partial}{\partial x} \left[ \frac{\partial T}{\partial x} + \lambda_x \frac{\partial T}{\partial z} \right] + \frac{\partial}{\partial z} \left[ \lambda_z \frac{\partial T}{\partial x} + \lambda_z \frac{\partial T}{\partial z} \right] - q_x \rho_f c_f \frac{\partial T}{\partial x} - q_z \rho_f c_f \frac{\partial T}{\partial z} = \left[ \phi \rho_f c_f + (1 - \phi) \rho_s c_s \right] \frac{\partial T}{\partial t}
\]

Fluid-Rock Isotope Transport:

\[
\frac{\partial}{\partial x} \left[ D_{xx} \frac{\partial R_{f}^i}{\partial x} + D_{xz} \frac{\partial R_{f}^i}{\partial z} \right] + \frac{\partial}{\partial z} \left[ D_{2x} \frac{\partial R_{i}^f}{\partial x} + D_{zz} \frac{\partial R_{i}^f}{\partial z} \right] - v_x \frac{\partial R_{i}^f}{\partial x} - v_z \frac{\partial R_{i}^f}{\partial z}
\]

\[
= \frac{\partial R_{i}^f}{\partial t} + \frac{\partial R_{f}^i}{\partial t} \ X_{rk} \ X_{f}
\]

Kinetic Fluid Rock Isotope Exchange:

\[
\frac{\partial R_{rk}}{\partial t} = A_{rk} \left[ \alpha_{rk} R_{f}^i - R_{rk} \right]
\]

Fluid-Rock Isotope Fractionation Factor:

\[
\alpha_{rk} = \frac{1}{\sum_{m=1}^{M} f_m \alpha_m}
\]

Kinetic Isotope Exchange Rate:

\[
r_{rk} = \frac{1}{\sum_{m=1}^{M} A_0^m \exp \left( -\frac{E_0^m}{RT} \right)}
\]

Helium 4 Transport equation

\[
\frac{\partial}{\partial x} \left[ D_{xx} \frac{\partial C_{ile}}{\partial x} + D_{xz} \frac{\partial C_{ile}}{\partial z} \right] + \frac{\partial}{\partial z} \left[ D_{2x} \frac{\partial C_{ile}}{\partial x} + D_{zz} \frac{\partial C_{ile}}{\partial z} \right] - v_x \frac{\partial C_{ile}}{\partial x} - v_z \frac{\partial C_{ile}}{\partial z}
\]

\[
= \frac{\partial C_{ile}}{\partial t} + R_{ile}
\]
Helium 3 Transport equation

\[
\frac{\partial}{\partial x} \left[ D_{xx} \frac{\partial C_{He}}{\partial x} + D_{xz} \frac{\partial C_{He}}{\partial z} \right] + \frac{\partial}{\partial z} \left[ D_{zx} \frac{\partial C_{He}}{\partial x} + D_{zz} \frac{\partial C_{He}}{\partial z} \right] - v_x \frac{\partial C_{He}}{\partial x} - v_z \frac{\partial C_{He}}{\partial z} = \frac{\partial C_{He}}{\partial t}
\]  

(59)

List of symbols

- \(k_{xx}, k_{zz}, k_{xz}, k_{zx}\) – components of the permeability tensor in the x- and z directions
- \(g\) – gravity constant
- \(\mu_f\) – fluid viscosity
- \(\mu_r\) – relative viscosity \((\mu_r = \frac{\mu_0}{\mu_f})\)
- \(h\) – Hydraulic head (m)
- \(\rho_0\) – Reference fluid density
- \(\rho_r\) – Relative density \([\rho_r = \frac{\rho_f - \rho_0}{\rho_0}]\)
- \(\rho_f\) – Temperature dependent fluid density
- \(x, z\) – Spatial coordinates
- \(\psi\) – stream function
- \(|k| = (k_{xx}k_{zz}-k_{xz}k_{zx})\)
- \(T\) – temperature
- \(\lambda_{xx}, \lambda_{zx}, \lambda_{xz}, \lambda_{zz}\) are the components of the thermal conductivity-dispersion tensor of the porous medium
- \(D_{xx}, D_{zx}, D_{xz}, D_{zz}\) are the components of dispersion-diffusion
- \(R_i^f\) – fluid isotopic ratio of species \(i\)
- \(R_i^r\) – bulk rock isotopic ratio
- \(t\) – time
- \(X_{mk}\) – fractional abundance of oxygen in the bulk rock phase
- \(\lambda_{mk}\) – fractional abundance of isotope in fluid
- \(A\) – bulk rock surface area (m²/mole)
- \(r_{mk}\) – bulk rock reaction rate (moles/m²·sec)
- \(\alpha_{mk}\) – bulk fluid rock equilibrium isotope exchange factor
- \(\alpha_m\) – temperature dependent fluid-rock isotope exchange factor
- \(M\) – total number of oxygen bearing mineral phases for a given rock
- \(A_0^m\) is the pre-exponential factor of the \(m\)th mineral phase (moles/m²·s)
- \(E_0^m\) is the activation energy for the exchange reaction (kcal/mole)
- \(R\) is the ideal law constant (kcal/mole·°K)
- \(T\) is temperature (°K)
- \(C_{4He}\) and \(C_{3He}\) are the helium-3 and helium-4 concentrations respectively
- \(R_{He}\) is the constant in situ production rate of helium-4

REFERENCES


Holk, G.J., Taylor, H.P., 2000, Water as a petrologic catalyst driving \(^{18}O/^{16}O\) homogenization and anatexis of the middle crust in the metamorphic core complexes of British Columbia, in Ernst,


Mulch, A., 2004, Integrated high-spatial resolution 40Ar/39Ar geochronology, stable isotope geochemistry, and structural analysis of extensional detachment systems: Case studies from the Prosgrunn-Kristiansand Shear Zone (S Norway) and the Shuswap Metamorphic Core Complex (Canada) [Ph.D. thesis]: Lausanne, Switzerland, University of Lausanne, 334 p.


