1
2
3 **The response of two-phase hydrothermal systems to changing magmatic heat**
4 **input at mid-ocean ridges**
5

6 Jaewoon Choi¹ and Robert P. Lowell

7 Department of Geosciences, Virginia Tech

8 Blacksburg, VA 24061

9 ¹Now at Korea Gas Company, 134-4 Sinseo-dong, Dong-gu, Daegu Korea 701-867
10

11 Submitted to: *Deep Sea Research II-Rona Volume*

12 July 30, 2014
13

14 Revised January 23, 2015
15

ABSTRACT

We develop numerical simulations of two-phase flow in a NaCl-H₂O fluid subject to time varying basal heat flux boundary conditions in order to understand the response of hydrothermal vent temperature and salinity to changing magmatic heat input. The results show that periodic changes in basal heat input on a time scale of several years will not be detected in a continuous time series record of temperature measurements. Fluctuations in vent salinity may be recorded, however. For models with monotonic decay of the magmatic heat flux, a decline in vent temperature may not be observed for several years; however, once single phase conditions are established at the base of the system, a pulse of brine-derived fluid is expected to appear at the surface, followed by a gradual decline of salinity to the seawater value. The pulse of brine salinity is expected to occur before an observed decline in vent temperature. Observed rapid changes in vent temperature and salinity associated with either eruptive or non-eruptive magmatic events are not likely a result of changes in basal heat flux.

1. Introduction

The linkage between the magmatic heat source and the overlying hydrothermal system at mid-ocean ridges is a subject of considerable scientific interest. As oceanic plates spread apart, upwelling mantle beneath the ridge undergoes decompressional melting, and melt is transported to a narrow zone of crustal formation. Seismic data shows that a zone of partially molten crust underlies a thin, mostly liquid lens of magma situated approximately 1 to 3 km beneath the seafloor (Detrick et al., 1987; Kent et al., 1990; Collier and Sinha, 1992; Singh et al., 2006; Jacobs et al., 2007; Van Ark et al., 2007; Carbotte et al. 2013). Heat transfer from the sub-axial magma lens (AMC) by conduction across a thin impermeable boundary layer drives the overlying hydrothermal system that discharges fluids with temperatures of ≈ 300 to 400°C (e.g., Cann and Strens, 1982; Lowell and Germanovich, 1994, 2004). Replenishment of the magma lens, commonly called the AMC (axial magma chamber), may result in diking events and magmatic eruptions (Germanovich et al., 2011). Moreover, Liu and Lowell [2009] have shown that relatively frequent magma replenishment is needed to maintain the observed temperature and heat output on decadal time scales. A key question concerns the response of the hydrothermal system to changes in magmatic heat input. Hydrothermal responses to other perturbations such as earthquakes, diking events, and changes in permeability have been addressed elsewhere (e.g., Germanovich et al., 2001; Wilcock, 2004; Ramondenc et al., 2008; 2011; Singh and Lowell, 2014)

To investigate mass, heat, and solute transport in a mid-ocean ridge hydrothermal system, seawater is generally assumed to be equivalent to a 3.2 wt% NaCl-H₂O solution (Bischoff and Rosenbauer, 1984). At the pressure and temperature conditions encountered near the top of the AMC, such a fluid would typically be in the two-phase, liquid-vapor region of the NaCl-H₂O

phase diagram. Figure 1 shows the NaCl-H₂O phase diagram in P - T - X space. The highest point of the two-phase liquid-vapor region as a function of P - T - X is shown by the solid red line called the critical isotherm. This curve connects the locus of critical points, which change with P - T - X conditions. For seawater the critical point is at approximately 407°C and 298 bars (Bischoff and Rosenbauer, 1988). The liquid-vapor equilibrium region lies beneath an isotherm below the critical point, and the salinity of the liquid and vapor in equilibrium at a given P - T are given by a horizontal tie line that intersects the vapor and liquid boundaries of the isotherm. High salinity liquid (commonly called brine) has a high density compared to bulk density, and thus it tends to remain at depth near the top of the AMC. Low salinity vapor is buoyant and tends to rise. This process is called phase separation. As low salinity vapor mixes with non-phase separated fluid during its ascent to the seafloor, it results in a fluid venting at the seafloor with salinity less than seawater. If some of the high salinity brine mixes with non-phase separated seawater and rises, the salinity of the resulting fluid exiting the seafloor is greater than seawater. Consequently, the salinity of vent fluids may vary with time and episodically vent vapor-derived and brine-derived fluids. A high-density layer at the base of hydrothermal system may act as a barrier between the overlying hydrothermal system and the magma chamber. In addition, diking events, whether eruptive or not, may result in phase separation in regions above the AMC. The salinity of a vent fluid phase separating at a shallow depth below the seafloor may be low shortly after a diking event, to be followed by a higher salinity fluid (Butterfield et al., 1997).

Models of two-phase flow in NaCl-H₂O fluids have only been developed recently (e.g., Bai et al., 2003; Kawada et al., 2004; Lewis and Lowell, 2004, 2009a,b; Driesner and Geiger, 2007; Coumou et al., 2008a,b, 2009a,b; Han et al., 2013; Singh et al., 2013). These simulations assumed fixed temperature or constant heat flux at the base of hydrothermal system. The fixed

temperature condition assumes that magma serves as an infinite reservoir of heat and hydrothermal heat transport is controlled by the vigor of the hydrothermal system, which is characterized by the Rayleigh number (e.g., Bejan, 1995; Lowell and Germanovich, 2004]. Hydrothermal heat output in the presence of a constant heat flux condition, however, is controlled by the rate at which the heat is conducted from the magma body (e.g., Lowell and Germanovich, 1994, 2004]. Neither of these boundary conditions is entirely realistic because heat transfer from a convecting magma body initially at its liquidus will result in cooling and crystallization leading to a decrease in heat output as a function of time (e.g., Cann and Strens, 1982; Liu and Lowell, 2009). Although Liu and Lowell (2009) considered heat transfer from a convecting, crystallizing, replenished magma chamber to the hydrothermal system, two-phase hydrothermal flow was not considered, and the hydrothermal system was assumed to respond instantly to changes in magmatic heat input. The principal objective of this study is to investigate the link between time dependent magmatic heat transfer and the evolution of vent temperature and salinity in a two-phase seafloor hydrothermal system.

In the next section of this paper, we present the mathematical formulation of the problem and discuss the basic features of the FISHERS code. The results for a variety of time varying basal heat flux distributions are given in section 3, and the implications of these results are discussed in section 4. Section 5 lists the main conclusions of this work.

2. Mathematical Formulation

2.1. The FISHERS code

FISHERS (Fully Implicit Seafloor Hydrothermal Event Simulator) is a FORTRAN code that uses the finite control volume method (Patankar, 1980) to solve the governing nonlinear

partial differential equations for conservation of mass, momentum, energy, and salt in a Darcian porous medium in a NaCl-H₂O hydrothermal system (Lewis, 2007).

The mass continuity equation is written as

$$\frac{\partial \phi \rho}{\partial t} + \nabla \cdot (\rho_v \vec{v}_v + \rho_l \vec{v}_l) = 0 \quad (1)$$

where ϕ is the porosity, ρ is bulk density, $\vec{v}$ is the Darcian velocity, and t is the time; subscripts v and l refer to the vapor and liquid phases, respectively. Table 1 provides a list of symbols and values of pertinent parameters. Darcy's Law is written for the vapor and liquid phases as

$$\vec{v}_v = -\frac{kk_{rv}}{\mu_v}(\nabla P - \rho_v \vec{g}) \quad (2)$$

$$\vec{v}_l = -\frac{kk_{rl}}{\mu_l}(\nabla P - \rho_l \vec{g}) \quad (3)$$

where k is permeability, k_r is the relative permeability, P is the pressure, μ is the dynamic viscosity, and $\vec{g}$ is the gravitational acceleration, respectively. FISHERS describes the relative permeabilities for the vapor and liquid phases using a linear relation for the volume saturations of vapor and liquid, respectively, with residual saturations set equal to zero. The bulk density ρ is given by

$$\rho = S_v \rho_v + S_l \rho_l \quad (4)$$

where S_l and S_v are the volume saturations of liquid and vapor, respectively. Conservation of energy is described by

$$\frac{\partial}{\partial t}[\phi\rho h + (1-\phi)\rho_r c_r T] + \nabla \cdot (\rho_v h_v \vec{v}_v + \rho_l h_l \vec{v}_l) = \nabla \cdot \lambda_m \nabla T \quad (5)$$

114 where h is the specific enthalpy, T is the temperature, and λ_m is the effective thermal
 115 conductivity. The subscript r refers to the rock; ρ and h without subscripts refer to bulk
 116 quantities. The bulk enthalpy is given by

$$117 \quad h = \frac{1}{f} (S_v r_v h_v + S_l r_l h_l) \quad (6)$$

118 Finally, we write conservation of salt as

$$119 \quad \frac{\partial}{\partial t}(\phi\rho X) + \nabla \cdot (\rho_v \vec{v}_v X_v + \rho_l \vec{v}_l X_l) = \nabla \cdot (\phi\rho_v D \nabla X_v + \phi\rho_l D \nabla X_l) \quad (7)$$

120 where X is the bulk salinity and D is the diffusivity of salt. The bulk salinity is given by

$$121 \quad X = \frac{1}{f} (S_v r_v X_v + S_l r_l X_l) \quad (8)$$

122 Note that, if porosity is constant, dividing by ϕ in equation (7) shows that salt is transported with the
 123 interstitial velocity $\vec{v} / \phi$ rather than the Darcian velocity.

124 Lewis (2007) formulates the thermodynamic equations of state in terms of lookup tables
 125 and an interpolation scheme. To construct these tables, the bulk densities and salinities between
 126 300°C and 800°C derive from Anderko and Pitzer (1993), whereas specific enthalpies in this
 127 temperature range are calculated using the framework of Tanger and Pitzer [1989]. Density and
 128 specific enthalpy values between 0°C and 300°C derive from Archer [1992]. Because FISHES is

used to simulate seafloor hydrothermal systems, the equations of state for the system cover the pressure range between 8.5-100 MPa, a temperature range between 0- 800°C and a salinity range between 0 and 100 wt% (Lewis, 2007). The code FISHERS is available online at <http://www.geophys.geos.vt.edu/rllowell/kaylal/>, and there is a user's guide with more details on the code. Lewis and Lowell (2009a,b), Steele-MacInnis et al. (2012a,b), Han et al., (2013), and Singh et al. (2013) present previous applications of FISHERS.

2.2 System geometry, boundary and initial conditions

To investigate the evolution of surface temperature and salinity as a function of time-varying heat flux at the base of the hydrothermal system, we considered a two-dimensional rectangular box that is 1.5 km deep and 4 km long with homogeneous permeability of 10^{-13} m^2 (Figure 2). This depth is consistent with seismic reflection data showing that the sub-axial magma lens (AMC) typically lies between approximately 1 and 3 km beneath the seafloor (Detrick et al., 1987; Singh et al., 1999, 2006; Canales et al., 2006; Van Ark et al., 2007; Carbotte et al., 2013). The permeability value chosen for the model is consistent with estimates from a range of seafloor hydrothermal systems for which heat output data are available (e.g., Lowell et al., 2013).

To save computation time, we used a half-model in the lateral dimension as shown by the vertical dashed line in Figure 2. Temperature and pressure at the top boundary correspond to nominal seafloor conditions of 10°C, 25MPa, respectively. Upstream conditions are applied at the top, so temperature and salinity of fluids exiting the surface are set equal to that at one node below the top boundary, whereas fluids entering the top of the system are given seawater salinity and surface temperature values. Impermeable, insulated conditions are imposed on the left and right hand boundaries. The bottom boundary is also impermeable. For these simulations, each

cell had dimensions of 25 m x 25 m, resulting in a total of 4800 cells. We have found from previous simulations using FISHES (e.g., Han et al., 2013; Singh et al., 2013) that this cell size provides adequate resolution of the temperature, salinity, and velocity structure within the computational domain. The principal difference between the simulations presented here and previous ones is in the use of time-dependent heat flux condition on the lower boundary rather than a constant temperature condition.

To simulate time-varying heat flux from the AMC, we considered a variety of basal boundary conditions: (1) a sinusoidal heat flux with a period of 6 years and an amplitude ranging between 150 and 50 W/m²; (2) a step function, random function, and exponential function between 200 and 15 W/m²; and (3) an analytical function of temporally decaying heat flux resulting from a simulated cooling, crystallizing magmatic sill. To represent the AMC, we assumed a high heat flux region of 500 m in length (e.g., Detrick et al., 1987; Van Ark et al., 2007; Carbotte et al., 2013) (Figure 2), and assumed that the heat flux decreased linearly to 1 W/m² at the right hand edge of the system. The 2D simulations approximate an across-axis circulation system, with a fairly broad zone of recharge compared to the narrower zone of discharge over the high heat flux region. This scenario may be representative of circulation at the Main Endeavour Field where deep recharge appears to be primarily off-axis (Johnson et al., 2000). Alternatively these simulations can approximate an along-axis circulation pattern in which the recharge occurs in a gap in the AMC along axis as suggested for the East Pacific Rise (EPR) near 9°50'N (Tolstoy et al., 2008).

Initially, we assume the temperature at the base is 450°C along the 500 m wide hot boundary and decreases linearly to 300°C at the right hand edge. We then assume that the initial temperature distribution increases linearly with depth from 10°C at the top until it matches the

temperature distribution along the bottom boundary. Initially, seawater salinity of 3.2 wt% was imposed throughout the model region, and the initial pressure distribution is hydrostatic.

2.3 Simulation method-the starting state

After the above set of initial conditions was imposed, we replaced the bottom temperature distribution with a periodic heat flux in a 500 m wide zone on the left side of the model domain with the expression

$$H = 50 + 25\sin 2\pi t \quad (9)$$

where H is in units of W/m^2 . Along the rest of the bottom boundary, we assumed the heat flux decreased linearly from 50 W/m^2 at 500 m to 1 W/m^2 at the right hand corner. This initial simulation was run for 150 years to establish a stable circulation pattern in the model domain. With the basal heat flux given by equation (6), phase separation does not occur, and single-phase hydrothermal flow takes place. After this initial simulation, the maximum basal temperature was 383°C and the maximum plume temperature was 311°C , respectively. We then used the PICKUP feature of FISHERS, which allows us to use the results of this initial simulation as the initial condition for all subsequent simulations.

To display the results of the simulations, we plot the basal temperature and surface temperature as well as the basal brine salinity and surface salinity as a function of time. The basal conditions provide insight into phase separation and the response of the system to the changing heat input. We plot the surface temperature and salinity to show how seafloor observables respond to the changing conditions at depth. We do not record the heat output of the system as a function of time, in part because there are so few heat flow data available for mid-ocean ridge hydrothermal systems for comparison. Moreover, because the simulations are time

dependent, the heat output is not expected to correlate with the heat input in a predictable manner.

3. Results

3.1 Constant heat flux

We assumed the heat flux from a simulated AMC with a half-width of 500 m to be 130W/m^2 , and to decrease linearly from 100 to 1 W/m^2 over the rest of the bottom boundary. This heat flux is sufficient to induce phase separation near the base of the convection system, and the convection system results in a number of high-temperature plumes situated above the AMC. Figure 3A shows the circulation pattern after 50 years in which a single high-temperature plume at depth breaks into two plumes near the surface. The maximum temperature of the plume is greater than $400\text{ }^\circ\text{C}$, and a small region of phase separation exists in the lower left-hand corner (Figure 3B). Broad red arrows in Figure 3A highlight the general circulation pattern.

Figures 4A, B show the bottom temperature and brine salinity, respectively, as a function of time at a number of specific distances from the left-hand corner boundary of the convecting regime. The plots show (red lines) that phase separation occurs near the left-hand bottom corner as temperature ranges between approximately 450°C and 500°C throughout the simulation, and a brine ranging between approximately 30 and 50 wt % salinity forms. The temperature and salinity fluctuations in space and time in these figures reflect the unsteady character of high Rayleigh number convection in these simulations (e.g., Cherkaoui and Wilcock, 1999; Coumou et al., 2009a; Perez et al., 2013). Farther from the bottom left-hand corner, temperature drops below the P - T conditions for phase separation, except for occasional fluctuations shown by the green line at 125 m from the corner. The salinity at 700 m remains at the seawater salinity of 3.2 wt %.

Figures 4C, D show the seafloor temperature and salinity, respectively, as a function of time at selected locations from the left-hand boundary. Seafloor temperatures range between 200°C and 350°C in plumes up to 475 m from the boundary, but at a distance of 700 m, the surface temperature of 10°C indicates fluid recharge occurs to the right of the high-temperature region. The temperature of these plumes places them in the single phase regime, but the salinity plot in Figure 4D shows the evidence of phase separation at depth as the seafloor salinity fluctuates about the seawater value shown by the horizontal dashed line. The salinity at the top left-hand corner fluctuates between 2.6~4.2 wt% NaCl, indicating that both vapor- and brine-derived fluids ascend from the bottom of the system. The pressure near the base of the system is approximately 38 MPa, and the mean temperature in the left hand corner is approximately 475°C. Under these conditions the brine and vapor salinities in equilibrium are 41.94 wt % and 0.34 wt %, respectively (Bischoff and Pitzer, 1989). The expected brine salinity is in accord with that shown in Figure 4B. The relatively high vapor- and brine-derived salinities observed on the seafloor reflect mixing with liquid phase seawater during ascent.

The fluctuating character of the temperature and salinity during the simulation is a reflection of the relatively high Rayleigh number of the convecting system. This value can only be assessed approximately for a phase separating system. By writing

$$Ra = \frac{\alpha g k H d^2}{\Gamma \alpha^* n} \quad (10)$$

and using parameter values from Table 1, one obtains $Ra \sim 1.4 \times 10^4$. For similar values of Ra in single-phase simulations, one often finds fluctuations on a decadal time scale (e.g., Coumou et al., 2009a).

Although the density of phase-separated brine is denser than seawater, the results show that some brine-derived fluid episodically enters the ascending fluid and exits the seafloor. This occurs in part by diffusion of the brine upward where it is then entrained into the ascending plume. It may also result in part from assuming the relative permeability is a simple linear function of the volume fractions of brine and vapor. If brine were assumed to be immobile below a certain volume fraction, brine might be better retained at depth.

3.2 Sinusoidal heat flux

In this simulation, the heat input from the magma chamber for a distance of 500 m along the base is assumed to be a sinusoid of the form

$$H(t)=100+50\sin\omega t \quad (11)$$

where $H(t)$ is in W/m^2 , ω is an angular frequency, $\omega = 2\pi/T$, and T is the period of the oscillation. Figures 5A, B show the bottom temperature and brine salinity, respectively, as a function of time at a number of points along the base of the system for $T = 2\pi$ year. After approximately 30 years of simulation, the temperature in the bottom left-hand corner enters the two-phase regime and begins to oscillate with the period of the basal heat flow. Temperatures farther from the corner remain in the liquid phase regime, and the oscillations suggest a periodicity corresponding to that of the basal heat flow superimposed on a longer periodicity that is similar to that observed in the case of constant basal heat flux (Figure 4A). Similarly, bottom salinity oscillates with a period corresponding to the basal heat flux. The mean brine salinity at the base of the system between 35 and 40 wt% is consistent with phase separation at a temperature of approximately 460°C and a pressure of 38 MPa. Figures 5C, D show the seafloor temperature and salinity, respectively, as a function of time at a few points within the discharge

plumes and one point of recharge 700 from the upper left-hand corner of the model. The 2π year period of the basal heat flow oscillation is not manifested in the seafloor temperature; however, oscillations in surface salinity are apparent (Figures 5C, D). The temperature perturbations are dampened both as a result of mixing between the hydrothermal fluid rising from the base of the system and heated single-phase seawater and also because the rising fluid thermally equilibrates with the surrounding rock as it traverses the system. On the other hand, salt is not exchanged between fluid and rock and moves with the interstitial velocity rather than with the Darcian velocity. Seafloor salinity fluctuates between 2.9~3.5 wt% NaCl, which is consistent with mixing between liquid seawater and the vapor or brine that phase separates at depth. Although seafloor salinity records the effect of periodic heat flux oscillations at the base of the system, the amplitude of the salinity oscillations at the surface is much smaller than at the base. A close comparison of Figures 5B, D shows that the left and edge of the system, the surface salinity is approximately 180° out of phase with the basal salinity at year 40. The phase difference changes with time because the basal salinity has a period of 2π , whereas the periodicity of the surface salinity is approximately 14% longer.

3.3 Periodic rectangular heat flux

To further explore the effect of periodic heat input at the base of the system, we consider periodic rectangular heat wave of the form:

$$H(t) = \begin{cases} 150 \text{ W/m}^2, & (n-1)T < t \leq \left(n - \frac{1}{2}\right)T \\ 50 \text{ W/m}^2, & \left(n - \frac{1}{2}\right)T < t \leq nT \end{cases} \quad (12)$$

where $T = 20$ years and $n=1, 2, 3, 4, 5$. We placed this heat flux along 500 m of the base, starting from the left-hand corner. Figure 6A plots the bottom temperature as a function of time

at a number of points along the base of the model domain. After approximately 4 years of simulation time, the temperature at the bottom left-hand corner (red line) oscillates about a mean temperature near 460°C, which is in the two-phase regime. The period of the oscillation is approximately 20 years, which is consistent with the periodic rectangular heat input wave emplaced in that region. Other plumes generally have temperatures in the liquid phase regime and show more irregular oscillations. Figure 6B depicts the brine salinity in the bottom left-hand corner oscillating between approximately 25 and 45 wt %, which is consistent with expected brine salinities at a pressure of around 38 MPa and the observed temperature oscillations at that location. The brine salinity of the two-phase flow approximately follows the 20-year oscillation period. Figure 6C displays the seafloor temperature and, as in the case of the shorter 6-year period sinusoidal heat input in Figure 5C, the plume temperatures at the seafloor do not record the 20-year periodic heat input at the base of the system. In Figure 6D, the seafloor salinity fluctuates between 2.9~3.5 wt%. The 20-year period of the basal heat flux is discernable in the salinity oscillations even though higher frequency perturbations are superimposed. A comparison of Figures 6C and Figure 6D indicates that the main salinity minima at the surface occur near the times of the salinity maxima at the base of the system. This phase difference is a result of the time needed for the fluid to ascend from the base of the system where phase separation takes place to the seafloor. As in the other simulations, the range of salinity oscillations is smaller at the seafloor than at the base of the system, which indicates that vapor- and brine-derived fluids ascending from the base of the domain are mixed with single-phase seawater.

3.4 Heat flux from a convecting, crystallizing, replenished magma chamber

Liu and Lowell (2009) showed that to maintain observed hydrothermal heat output between ~ 100 and 500 MW on decadal time scales, magma replenishment to the AMC at rates

between $\sim 10^{-7}$ and 10^{-8} m/s are required. In their modeling, however, they linked hydrothermal heat output and vent temperature instantaneously to the rate of heat transfer from the AMC. Liu and Lowell (2009) also neglected the effects of phase separation. To better understand the response of a phase separating hydrothermal system to changing heat output from a convecting, crystallizing, replenished AMC, we incorporated the time varying heat flux from Liu and Lowell (2009) as a basal boundary condition on FISHES. We determined the form of the boundary condition by constructing an analytical function to fit the heat output profile and dividing the heat output by the assumed area of the AMC to obtain a heat flux as a function of time. The results shown in Figure 7, for the situations of no magma replenishment and magma replenishment at a rate of 1.8×10^{-8} m/s, were then used as the basal boundary condition in FISHES simulations. These heat flux curves assume a magma convection model in which crystals descend rapidly to the floor of the AMC and in the case of magma replenishment leads to a thickening AMC with constant surface area (Liu and Lowell, 2009).

3.4.1 No replenishment

Figure 8A plots the bottom temperature versus time at selected locations in the case of magmatic heat flux without magma replenishment. Because the heat flux at time zero is much higher than is realistic, the FISHES simulations start when the heat flux has declined to the value at year 10 (Figure 7). This simulation results in a period of two-phase flow for approximately 18 years, followed by a sharp decrease in bottom temperature into the single-phase region (red line in Figure 8A) as the basal heat flux decays. In Figure 8B a corresponding increase in brine salinity occurs at the base of the system, followed by a rapid return to seawater salinity values coincident with the decay of the basal temperature.

Figure 8C indicates that surface vent temperatures remain relatively high long after the basal temperature has dropped into the single-phase regime. Hence, it is not easy to correlate vent temperature to the rapidly decaying heat flux at the base of the system. Figure 8D shows that seafloor salinity fluctuates between approximately 2.8 and 3.4 wt% during the period of two-phase flow. These fluctuations, which are similar to those observed in previous figures, are followed by a sharp increase in salinity within ~ one year after phase separation ceases at the base of the system. This rise in vent salinity, followed by a return to the seawater value of 3.2 wt%, is a result of flushing brine from the base of the system and mixing it with seawater. In this simulation, the stored brine is removed in less than a decade.

3.4.2 Magma replenishment.

When magma replenishment is considered, the simulation begins with a heat flux of approximately 200 W/m² (Figure 7). The heat flux decays in time as magma replenishment at a rate of 1.8×10^{-8} m/s stabilizes the heat flux to a final value of 100 W/m² after approximately 30 years (Figure 7). This value of basal heat flux results in a single-phase hydrothermal system. Figure 9A depicts the bottom temperature versus time at selected locations. After 3 years, the temperature of the bottom left corner (red line) reaches 500°C, decreases for a time to approximately 425°C, and then settles to slightly more than 450°C until the last 10 years. During the last 10 years, the temperature drops into the single-phase regime as a result of the stable, relatively low heat flux at the base. The bottom brine salinity (red line in Figure 9B) follows the basal temperature indicating phase separation near the base of the system. At approximately 90 years, the peak bottom temperature lies in the single-phase regime, and the bottom salinity returns to the seawater value.

In Figure 9C the surface temperature distribution looks similar to the case without replenishment (Figure 8C), reflecting the muted response of seafloor temperature to conditions at depth. Seafloor salinity (Figure 9B) fluctuates between 2.6~4.0 wt% NaCl, which is consistent with phase-separated fluid ascending from the bottom of the system and mixing with seawater. This pattern is similar as constant heat flux case (Figure 4D) until the heat flux decreases and the fluid enters the liquid phase regime. The final spike in vent salinity at approximately 95 years into the simulation reflects the flushing of brine stored at the base of the system.

4. Discussion and Conclusions

Temperature and salinity of hydrothermal vents are two key measurable parameters that provide insight into the physical and chemical processes occurring at depth in a mid-ocean ridge hydrothermal system. Temperature variations may reflect the cooling or rejuvenation of the magmatic heat source, permeability changes, water-rock reactions, and biogeochemical reactions (Scheirer et al., 2006). Vent fluid salinity and its variations reflect the occurrence of phase separation and hence provide insight into temperature and pressure conditions at depth. The resulting salinity of hydrothermal vents is affected by mixing processes, which may be affected by the permeability distribution in the crust. In high Rayleigh number convection regimes that likely occur at mid-ocean ridges, some variations may result from the unsteady nature of the convection regime. Volcanic eruptions, dike intrusion, and earthquakes can also cause temperature and salinity variations by changing both temperature and permeability within the circulation system. Hence, it is often difficult to infer how changes in temperature and salinity at the seafloor relate to changes in subsurface conditions.

4.1 Summary of results

In this paper, we have investigated the link between time-dependent magmatic heat transfer and two-phase hydrothermal circulation at a mid-ocean ridge axis in order to determine how temperature and salinity at the seafloor would respond to a variety of time-varying heat input functions and to determine whether the observed changes in vent fluid parameters could be used to infer temporal changes in magmatic heat input. The principal findings from the simulations presented here are enumerated below.

1. Bottom temperatures in regions where phase separation occurs respond rapidly to periodic changes in heat flux. In the two-phase region near the bottom, the pressure remains nearly constant, and salinity changes to be in equilibrium with the P-T conditions at any given time. For periodic heat inputs, bottom temperature changes in the two phase region fluctuate by approximately $\pm 25^{\circ}\text{C}$ about the mean of approximately 460°C , leading to changes in basal brine salinity between 25 and 45 wt % NaCl (e.g., see Figures 5B and 6B). Corresponding vapor salinities are not shown, but range between approximately 0.3 and 0.8 wt % NaCl (Bischoff and Pitzer, 1989).

2. Seafloor temperatures do not respond in a clear way to changes in bottom heat flux. This appears to be true whether the heat flux changes occur periodically on time scales ranging between several years to decades, or whether there is a relatively rapid monotonic decline in heat flux. As temperature changes at depth propagate upward, thermal equilibration with the adjacent rock and mixing between ascending, hot, phase-separated fluids and cooler liquid phase seawater severely dampen the response of the seafloor temperature. As heat is transferred with the Darcian velocity, which in these simulations is $\sim 5 \times 10^{-6}$ m/s, it would take approximately 10 years for heat to be transferred from the top of the AMC at a depth of 1.5 km to the seafloor. Hence even in the case of declining magmatic heat flux, the surface temperature response would lag the

declining basal temperatures by several years. Given the relatively short history of observation at most vent fields, declining magmatic heat input may be hard to detect from observations of vent temperature.

3. Seafloor salinities respond more clearly to changes in the thermal regime at depth. For periodic and square wave heat input, with periods of 2π and 20 years, respectively, vent fluid salinity fluctuates about the seawater value of 3.2 wt %. At the seafloor, the hydrothermal system is in the single-phase regime, and the vent fluid salinities reflect mixing between vapor or brine with seawater. At the high-Rayleigh numbers estimated for these simulations, temporal changes in these vapor- and brine-derived fluids occur on a time scale of decades. For the constant heat flux simulation shown in Figure 4, the seafloor salinity ranges between 2.6 and 4.2 wt%. Superimposed on these temporal changes, oscillations in seafloor salinity with nearly the same period as the heat flux variations are clearly discernible (Figures 5 and 6).

4. Seafloor salinities, whether responding to periodic heat fluxes at depth or to a monotonically decaying heat flux, exhibit a time delay relative to changes at depth. This delay mainly reflects the time needed for fluid to travel from the base of the system to the seafloor. Because the fluid does not exchange salt with the adjacent rock, as it does heat, salinity changes travel with the interstitial velocity rather than the Darcian velocity (compare equations (5) and (7)). Hence in the case of a decaying heat flux, salinity changes appear at the surface before temperature changes. The time delay of the last brine-derived fluid was about 5 years (Figure 8). As the heat flux decreases to where only single-phase flow occurs, the bottom salinity returns to the seawater value and the last pulse of brine-derived fluid is eventually flushed out of the system. Then, the salinity gradually returns to the seawater value throughout the entire convection domain.

The range in seafloor salinities produced in these model simulations is smaller than observed at the EPR and MEF hydrothermal systems (see Figures 10B and 11B below). This difference stems in part from the recognition that the lowest values of salinity measured at the seafloor occur in conjunction with eruptions or diking events such that phase separation occurs near the seafloor. In this case, phase separated fluid has less opportunity to mix with seawater.

4.2 Model simplifications

The results of the simulations presented here stem from a hydrothermal circulation model that assumes a constant, homogeneous permeability of 10^{-13} m^2 . This value of permeability is of the same order of magnitude that is calculated for the upflow zone as EPR 9°50' N and the MEF. As intact permeability of basaltic rocks is many orders of magnitude lower than this (e.g., Gleeson et al., 2011), the value used in the simulations corresponds to that of fractured basalt. It represents a mean value for the entire upflow zone, however, and the actual permeability is likely to be highly heterogeneous. The range and distribution of heterogeneity is unknown, however, so assuming a relatively high, but uniform permeability is a reasonable first-order approach.

The simulations also assume that Darcy's Law is valid, which means that the 25 m cell size is a representative equivalent volume for the fractured-porous medium and that thermal equilibrium between the fluid and the rock is maintained on that scale. The time needed for thermal equilibrium between the rock and fluid can be approximated by the conductive time scale $\tau \sim d^2/a^*$, where d is $\frac{1}{2}$ the crack spacing. For thermal equilibrium between the fluid and rock, the time for the fluid to flow along the crack must be less than τ . In the simulations presented here, the maximum interstitial velocity is approximately $5 \times 10^{-5} \text{ m/s}$, so the time to traverse 25 m is approximately $5 \times 10^5 \text{ s}$. Using this value for τ yields a value of $2d \sim 1.4 \text{ m}$. This

suggests that, if the crack spacing in the medium is on a scale of 1.4 m or less, then Darcy's Law will hold and thermal equilibrium between fluid and rock is maintained. This seems to be a reasonable assumption throughout most of the upflow zone based on crack characteristics in ophiolites (e.g., Nehlig, 1994; van Everdingen, 1995). In the upper crust, where upflow is focused into discrete vents, larger, through-going cracks may exhibit greater fluid velocities. In these situations, a temperature perturbation in the fluid may not maintain thermal equilibrium with the adjacent rock.

4.3 Application to the EPR and MEF hydrothermal systems

Time series observations at the EPR 9°50' N site and the Main Endeavour Field on the Juan de Fuca Ridge (Figures 10 and 11) show that vent temperatures and salinities exhibit significant changes during the decade-plus period of observation. The key question is whether the observed changes are related to changes in heat flux from the AMC as described by the models developed in this paper. The results of the simulations shown here indicate that the answer appears to be negative.

At EPR, temperature increases and low values of salinity occur near the time of eruptive events in 1991-1992 (Haymon et al, 1993; Rubin et al., 1994) and again in 2005-2006 (e.g., Tolstoy et al., 2006; Soule et al., 2007). In between these eruptions, short-lived temperature excursions of several degrees were recorded on HOBO probes at Bio9 in 1995 (Sohn et al., 1998) and 2003 (Tolstoy et al., 2008a). These temperature perturbations may be related to non-eruptive diking events (Germanovich et al., 2011), which would be consistent with relatively frequent magma replenishment as proposed by Liu and Lowell (2009). During the time period between 1991 and 2007, discrete temperatures measurements at Bio9, Bio9', and P exhibited

fluctuations of approximately $\pm 15^{\circ}\text{C}$ about a mean of approximately 375°C . It is difficult to correlate the temperature record shown in Figure 10A with subsurface magmatic processes because the hypothesized occurrence of the non-eruptive magmatic events cannot be confirmed, and the temperature record is sparse. Vent temperatures at TWP near the southern end of the hydrothermal system gradually declined between 1994 and 2002, and the vent became extinct in 2003. Io also became extinct by 2007. The decay of these vents may have resulted from cold water recharge occurring near these sites (Tolstoy et al., 2008b) rather than from declining heat flux from the AMC.

Figure 10B shows that vent salinities declined sharply around the time of the eruptive events of 1991-1992 and 2005-2006. Such a decline is expected because phase separation occurs at lower pressures, and the vapor-phase fluid may rise rapidly to the seafloor without significant mixing. Following these eruptive events, vent salinities rose, with peak salinities occurring some months following the non-eruptive event in 1995. Then after a long period of decline the vent salinities exhibited a small increase that peaked in 2004. If the non-eruptive events recorded by the HOBO temperature probes in 1995 and 2003 reflected changes in heat input into the hydrothermal system the models presented here indicate that seafloor temperature perturbation would be difficult to detect, but one might expect to observe salinity changes. Although one can observe quasi-periodic salinity changes in Figure 10B, there is not enough information to conclusively attribute these changes to changes in magmatic heat flux. Moreover, changes in crustal permeability could also result in perturbations to vent temperature and salinity (Singh and Lowell, 2015), and such changes are not incorporated into the models presented here.

An important feature of the MEF on the Juan de Fuca Ridge is the observed north to south gradient in temperature and salinity shown prior to 1999 in Figure 11A and B. Both

temperature and salinity show a marked response to a non-eruptive lateral dike event in 1999 (Johnson et al., 2000; Lilley et al., 2003; Bohnenstiehl et al. 2004; Figure 11), with salinities at Puffer and Sully in the southern part of the vent field exhibiting very low values. In the years since this event, vent temperatures in the southern MEF declined significantly while vent salinity increased well above their pre-event values. While the southern part of the MEF is waning (Kelley et al., 2012), seismicity data suggests that magma replenishment is occurring in the northern part (Wilcock et al., 2009).

The numerical simulation presented in Figure 8 for a constant permeability simulation with a decaying AMC heat flux suggests that a detectable decline in vent temperatures would not occur until some years after the bottom temperature declined, and the surface temperature decline would be gradual. Observable changes in vent salinity would occur within approximately one year following the return to single-phase conditions at depth. Although the time series is short, Figure 11 shows that vent temperature reached a peak approximately one year following the dike event, whereas low salinity vapor appeared more quickly. Vent temperatures then began to decay as salinity continued to rise. The temperature and salinity data collected between 1999 and 2005 may still reflect the impact of this event on heat and salt transport. By 2007, however, vent temperature at Sully was only 230°C (Kelley et al., 2012), a considerable decline from pre-event temperatures. The decay of the southern part of the MEF, while the northern part appears to be undergoing magma replenishment, suggests that these two parts of the MEF have been fed by different fluid circulation regimes (Kelley et al., 2012). From the model results in Figure 8, one might speculate that the decline in venting in the southern MEF following the dike event indicates that melt replenishment of the AMC in the southern part of the MEF decreased sometime before the dike event, and the laterally propagating dike perturbed the

hydrothermal system there. Fluid flow paths and permeability distributions beneath MEF are likely to be highly heterogeneous, however, (e.g., Singh et al., 2013), and these complexities are not accounted for in the simulations presented here.

In summary, the models developed here broadly suggest that temporal changes in basal magmatic heat flux will likely result in observable changes in vent salinity, but not in vent temperature. The uncertain and heterogeneous nature of crustal permeability, the possible temporal and spatial distribution of magmatic heat sources associated with eruptions and non-eruptive diking events, and the potential effect of these events on crustal permeability likely result in complex perturbations in vent temperature and salinity that are not accounted for in these simulations. Conversely, interpreting observed changes in vent temperature and salinity as a result of changes basal heat flux is highly problematical. In order to better understand how hydrothermal systems respond to magmatic perturbations, models involving temporal and spatial variations in permeability, and the impact of heterogeneous permeability distributions are needed along with better time series data on both temperature and salinity.

Acknowledgments

This work was supported in part by NSF Grant OCE 0926418 to RPL. The authors appreciate the comments of Shreya Singh and two anonymous reviewers on an earlier version of the manuscript.

References

- Anderko, A., Pitzer, K., 1993. Equation-of-state representation of phase equilibria and volumetric properties of the system NaCl-H₂O above 573K, *Geochim. Cosmochim. Acta* 57, 1657-1680.
- Archer, D., 1992. Thermodynamic properties of the NaCl+H₂O system II. Thermodynamic properties of NaCl (aq), NaCl·2H₂O (cr), and phase equilibria, *J. Phys. Chem. Ref. Data* 21, 793-820.
- Bai, W., Xu, W., and Lowell, R.P., 2003. The dynamics of submarine geothermal heat pipes. *Geophys. Res. Lett.* 30(3), 1108, doi:10.1029/2002GL016176.
- Bejan, A., 1995. *Convection heat transfer*, Wiley & Sons, New York.
- Bischoff, J., Rosenbauer, R., 1984. The critical point and two-phase boundary of seawater, 200° - 500 °C. *Earth Planet. Sci. Lett.* 68, 172-180.
- Bischoff, J., Rosenbauer, R., 1988. Liquid-vapor relations in the critical region of the system NaCl-H₂O from 380 to 415°C: A refined determination of the critical point and two-phase boundary of seawater, *Geochim. Cosmochim. Acta* 52, 2121-2126.
- Bischoff, J.L., Pitzer, K.S., 1989. Liquid-vapor relations for the system NaCl-H₂O: Summary of the P-T-x surface from 300° to 500°C. *Am. J. Sci.* 289, 217-248.
- Bohnenstiel, D.R., Dziak, R.P., Tolstoy, M., Fox, C.G., Fowler, M., 2004. Temporal and spatial history of the 1999-2000 Endeavour Segment seismic series, Juan de Fuca Ridge. *Geochim. Geophys. Geosyst.*, 5, Q09003, doi:10.1029/2004GC000735.
- Butterfield, D.A., McDuff, R.E., Mottl, M.J., Lilley, M.D., Lupton, J.E., Massoth, G.J., 1994. Gradients in the composition of hydrothermal fluids from the Endeavour segment vent field: phase separation and brine loss. *J. Geophys. Res.* 99, 9561-9583.
- Butterfield D.A., Jonasson, I.R., Massoth, G.J., Feely, R.A., Roel, K.K., Embley, R.E., Holden, J.F., McDuff, R.E., Lilley, M.D., Delaney, J.R., 1997. Seafloor eruptions and evolution of hydrothermal fluid chemistry. *Phil. Trans. R. Soc. A* 355, 369-386.
- Canales, J.P., Singh, S.C., Detrick, R.S., Carbotte, S.M., Harding, A., Kent, G.M., Diebold, J.B., Babcock, J., Nedimovic, M.R., 2006. Seismic evidence for variations in axial magma chamber properties along the southern Juan de Fuca Ridge. *Earth. Planet. Sci. Lett.* 246, 353-366.
- Cann, J., Strens, M. 1982. Black smokers fuelled by freezing magma. *Nature* 298, 147-149, doi:10.1038/298147a0.
- Carbotte, S.M., Marjanovic, M., Carton, H., Mutter, J.C., Canales, J.P., Nedimovic, M., Han, S., Perfit, M.R., 2013. Fine-scale segmentation of the crustal magma reservoir beneath the East Pacific Rise. *Nature Geosci.* 6, 866-870, doi:10.1038/NGE01933.

573
574 Cherkaoui, A.S.M., Wilcock, W.S.D., 1999. Characteristics of high Rayleigh number two-
575 dimensional convection in an open-top porous layer heated from below. *J. Fluid Mech.* 394, 241-
576 260.

577 Collier, J.S., Sinha, M.C., 1992, Seismic mapping of a magma chamber beneath the Valu Fa Ridge,
578 Lau Basin. *J. Geophys. Res.* 97, 14031-14053.

579 Coumou, D., Driesner, T., Geiger, S., Heinrich, C.A., Matthäi, S., 2006. The dynamics of mid-
580 ocean ridge hydrothermal systems: splitting plumes and fluctuating vent temperatures.,
581 *Earth Planet. Sci. Lett.* 245, 218-235.

582
583 Coumou, D., 2008. Numerical simulation of fluid flow in mid-ocean ridge hydrothermal
584 Systems. Ph.D. thesis, ETH Zurich, Zurich, Switzerland.

585
586 Coumou, D., Driesner, T., Heinrich, C., 2008a. Heat transport at boiling, near-critical conditions,
587 *Geofluids* 8, 208-215.

588
589 Coumou, D., Driesner, T., and Heinrich, C., 2008b. The structure and dynamics of mid-
590 ocean ridge hydrothermal systems. *Science* 321, 1825-1828.

591
592 Coumou, D., Driesner, T., Geiger, S., Paluszney, A., Heinrich, C.A., 2009a. High-resolution
593 three-dimensional simulations of mid-ocean ridge hydrothermal systems. *J. Geophys. Res.*, 114,
594 B07104, doi:10.1029/2008JB006121.

595
596 Coumou, D., Driesner, T., Weis, P., Heinrich, C., 2009b. Phase separation, brine formation, and
597 salinity variation at Black Smoker hydrothermal systems. *J. Geophys. Res.* 114, B03212,
598 doi:10.1029/2008JB005764.

599
600 Detrick, R. S., Buhl, P., Vera, E., Mutter, J., Orcutt, J., Madsen, J., Brocher, T., 1987.
601 Multichannel seismic imaging of a crustal magma chamber along the East Pacific Rise. *Nature*
602 326, 35-41.

603
604 Driesner T., Geiger, S., 2007. Numerical simulation of multiphase fluid flow in hydrothermal
605 systems. *Rev. Min. Geochem.* 65, 187-212.

606
607 Driesner, T., Heinrich, C. 2007. The system H₂O-NaCl. Part I: Correlation formulae for phase
608 relations in temperature-pressure-compositions space from 0 to 1000 °C, 0 to 5000 bar, and
609 0 to 1 XNaCl, *Geochim. Cosmochim. Acta* 71, 4880-4901.

610
611 Faust, C. R., Mercer, J.W., 1979. Geothermal reservoir simulation: 1. Mathematical models for
612 liquid- and vapor-dominated hydrothermal systems. *Water Resour. Res.* 15, 23-30.

613
614 Foustoukos, D. I., Pester, N.J, Ding, K., Seyfried Jr., W.E., 2009. Dissolved carbon species in
615 associated diffuse and focused flow hydrothermal vents at the Main Endeavour Field, Juan de
616 Fuca Ridge: Phase equilibria and kinetic constraints. *Geochem. Geophys. Geosyst.* 10, Q10003,
617 doi:10.1029/2009GC002472.

- Geiger, S., Driesner, T., Heinrich, C.A., Matthäi, S.K., 2006.a Multiphase thermohaline convection in the Earth's crust: I. A new finite element-finite volume solution technique combined with a new equation of state for NaCl-H₂O. *Transp. Porous Media* 63, 399-434.
- Geiger, S., Driesner, T., Heinrich, C.A. Matthäi, S.K., 2006b. Multiphase thermohaline convection in the Earth's crust: II. Benchmarking and application of a finite element-finite volume solution technique with a NaCl-H₂O equation of state. *Transp. Porous Media* 63, 435-461.
- Germanovich, L.N., Lowell, R.P., Ramondenc, P., 2011. Magmatic origin of hydrothermal response to earthquake swarms: constraints from heat flow and geochemical data at 9°50'N, East Pacific Rise. *J. Geophys. Res.* 116, B05103, [http:// dx.doi.org/10.1029/2009JB006588](http://dx.doi.org/10.1029/2009JB006588).
- Gleeson, T., Smith, L., Moosdorf, N., Hartman, J., Durr, H.H., Manning, A.H., van Beek, L.P.H., Jellinek, A.M., 2011. Mapping permeability over the surface of the Earth. *Geophys. Res. Lett.* 38, L02401, doi:10.1029/2010GL045565.
- Han, L., Lowell, R.P., Lewis, K.C., 2013. Dynamics of two-phase hydrothermal systems at a surface pressure of 25 MPa. *J. Geophys. Res.* 118, 2635-2647, doi:10.1002/JGRB.50158.
- Haymon, R. M., et. Al., 1993. Volcanic eruption of the mid-ocean ridge along the East Pacific Rise crest at 9°45'-52'N: direct submersible observations of seafloor phenomena associated with an eruption event in April, 1991. *Earth Planet. Sci. Lett.* 119, 85-101.
- Jacobs, A.M., Harding, A.J., Kent, G.M., 2007. Aclal crustal structure of the Lau back-arc basin from velocity modeling of multichannel seismic data., *Earth Planet. Sci. Lett.* 259, 239-255.
- Johnson, H.P., Hutnak, M., Dziak, R.P., Fox, C.G., Urcuyo, I., Cowan, J.P., Nabelek, J., Fisher, C., 2000. Earthquake-induced changes in a hydrothermal system on the Juan de Fuca mid-ocean ridge. *Nature* 407, 174-177.
- Kawada, Y., Yoshida, S., Watanabe, S.I., 2004. Numerical simulations of mid-ocean ridgehydrothermal circulation including the phase separation of seawater. *Earth Planets Space* 56, 193-215.
- Kelley, D.S. et al., 2012. Endeavour Segment of the Juan de Fuca Ridge, one of the most remarkable places on Earth, *Oceanography* 25, 44-61, <http://dx.doi.org/105670/oceanog.2012.03>.
- Kent, G.M., Harding, A.J., Orcutt, J.A. 1990. Evidence for a smaller magma chamber beneath the East Pacific Rise at 9°30'N. *Nature* 344, 650-652.
- Lewis, K.C., 2007. Numerical modeling of two-phase flow in the sodium chloride-water system with applications to seafloor hydrothermal systems. Ph.D. thesis, Georgia Institute of Technology, Atlanta.

- Lewis, K.C., Lowell, R.P., 2004. Mathematical modeling of phase separation of seawater near an igneous dike. *Geofluids* 4, 197-209.
- Lewis, K.C., Lowell, R.P., 2009a. Numerical modeling of two-phase flow in the NaCl-H₂O system: introduction of a numerical method and benchmarking. *J. Geophys. Res.* 114, B05202, doi:10.1029/2008JB006029.
- Lewis, K.C., Lowell, R.P., 2009b. Numerical modeling of two-phase flow in the NaCl-H₂O system: 2. Examples. *J. Geophys. Res.* 114, B08204, doi:10.1029/2008JB006030.
- Lilley, M.D., Butterfield, D.A., Lupton, J.E., Olson, E.J., 2003. Magmatic events produce rapid changes in hydrothermal vent chemistry. *Nature* 422, 878-881.
- Liu, L., Lowell, R.P. 2009. Models of hydrothermal heat output from a convecting, crystallizing, replenished magma chamber beneath an oceanic spreading center. *J. Geophys. Res.* 114, B02102, doi:10.1029/2008JB005846.
- Lowell, R.P., Farough, A., Hoover, J., Cummings, K.C. 2013. Characteristics of magma-driven hydrothermal systems at oceanic spreading centers. *Geochem. Geophys. Geosyst.* 14, doi:10.1002/ggge.20109.
- Lowell, R.P., Germanovich, L.N., 1994. On the temporal evolution of high-temperature hydrothermal systems at ocean ridge crests. *J. Geophys. Res.* 99, 565-575.
- Lowell, R. P., Germanovich, L.N., 2004. Hydrothermal processes at mid-ocean ridges: results from scale analysis and single-pass models. In: German, C., Lin, J., Parson, L.M. (Eds.), *Mid-Ocean Ridges: Hydrothermal Interactions between the Lithosphere and Oceans*. AGU, Washington, D.C., *Geophys. Monogr. Ser.* 148, 219-244.
- Nehlig, P., 1994. Fracture and permeability analysis in magma-hydrothermal transition zones in the Samail ophiolite (Oman). *J. Geophys. Res.* 99, 589-601.
- Patankar, S., 1980. *Numerical Heat Transfer and Heat Flow*. Taylor and Francis, London.
- Perez, F., Mügler, C., Jean-Baptiste, P., Charlou, J.L., 2013. Coupled modeling of thermics and hydrogeology with the Cast3M code: application to the Rainbow hydrothermal field (Mid-Atlantic Ridge, 36°14'N). *Comput. Geosci.* 17, 217-237, doi:10.1007/s10596-012-9327-x.
- Ramondenc, P., Germanovich, L.N., Lowell, R.P., 2008. Modeling the hydrothermal response to earthquakes with application to the 1995 event at 9°50' N, East Pacific Rise. In: Lowell, R.P., Seewald, J.S., Metaxas, A., Perfit, M.R. (Eds.), *Magma to Microbe: Modeling Hydrothermal Processes at Oceanic Spreading Centers*. AGU Washington, D.C., *Geophys. Monogr. Ser.* 178, 97-122.
- Rubin, K.H., Macdougall, J.D., Perfit, M.R., 1994. ²¹⁰Po-²¹⁰Pb dating of recent volcanic eruptions on the sea floor. *Nature* 368, 841-844.

- Scheirer, D. S., Shank, T.M., Fornari, D.J., 2006. Temperature variations at diffuse and focused flow hydrothermal vent sites along the northern East Pacific Rise. *Geochem. Geophys. Geosyst.* 7, Q03002, doi:10.1029/2005GC001094.
- Singh, S., Lowell, R.P., 2015. Thermal response of mid-ocean ridge hydrothermal systems to perturbations. *Deep Sea Research II* (this issue).
- Singh, S., Lowell, R.P., Lewis, K.C., 2013. Numerical modeling of phase separation at the Main Endeavour Field, Juan de Fuca Ridge., *Geochem. Geophys. Geosyst.* 14, doi:10.1002/ggge.20249.
- Singh, S.C., Collier, J.S., Harding, A.J., Kent, G.M., Orcutt, J.A., 1999. Seismic evidence for a hydrothermal layer above the solid roof of the axial magma chamber at the southern East Pacific Rise. *Geology* 27, 219-222.
- Singh, S.C., Crawford, W.C., Carton, H., Seher, T., Combier, V., Cannat, M., Canales, J.P., Düşünür, D., Escartin, J., Miranda, J.M., 2006. Discovery of a magma chamber and faults beneath a Mid-Atlantic Ridge hydrothermal field. *Nature* 332, 1029-1032.
- Sohn, R.A., Fornari, D.J., Von Damm, K.L., Hildebrand, J.A., Webb, S.C., 1998. Seismic and hydrothermal evidence for a cracking even on the East Pacific Rise crest at 9°50'N. *Nature* 396, 159-161.
- Soule, S.A., Fornari, D.J., Perfit, M.R., Rubin, K.H., 2007. New insights into mid-ocean ridge volcanic processes from the 2005-2006 eruption of the East Pacific Rise, 9°46'N-50°56' N. *Geology* 35, 1079-1082, doi: 10.1040/G23924A.1
- Steele-MacInnis, M., Han, L., Lowell, R.P., Rimsstidt, J.D., Bodnar, R.J., 2012a. The role of fluid phase immiscibility in quartz dissolution and precipitation in sub-seafloor hydrothermal systems. *Earth Planet. Sci. Lett.* 321-322, 139-151. doi:10.1016/j.epsl.2011.12.037.
- Steele-MacInnis, M., Han, L., Lowell, R.P., Rimsstidt, J.D., Bodnar, R.J., 2012b. Quartz precipitation and fluid inclusion characteristics in sub-seafloor hydrothermal systems associated with volcanogenic massive sulfide deposits. *Centr. Euro. J. Geosci.* 4, 275-286., doi: 10.2478/s13533-011-0053-z.
- Strens, M.R., Cann, J.R., 1982. A model of hydrothermal circulation in fault zones at mid ocean ridge crests. *Geophys. J. R. Astron. Soc.* 71, 225-240.
- Tanger, J., Pitzer, K., 1989. Thermodynamics of NaCl-H₂O: a new equation of state for the near-critical region and comparisons with other equations for adjoining regions., *Geochim. Cosmochim. Acta* 53, 973-987.
- Tolstoy, M., Waldhauser, F., Bohnenstiehl, D.R., Weekly, R.T., Kim, W.Y., 2008b. Seismic identification of along-axis hydrothermal flow on the East Pacific Rise. *Nature* 451, 181-184.

- Tolstoy, M. et al., 2006. A seafloor spreading event captured by seismometers. *Science* 314, 1920-1922, doi:10.1126/science.1133950.
- Tolstoy, M., Crone, T.J., Felix, W., Bohnenstiehl, D.R., Fornari, D.J., Von Damm, K., 2008b. Seismic activity associated with temperature perturbations at Bio 9 hydrothermal vent on the East Pacific Rise at 9°50'N. *Eos, Trans. AGU*, 89, Fall Meet. Suppl. Abstract B21A-0326.
- Van Ark, E. M., Detrick, R.S., Canales, J.P., Carbotte, S.M., Harding, A.J., Kent, G.M., Nedimovic, M.R., Wilcock, W.S.D., Diebold, J.B., Babcock, J.M., 2007. Seismic structure of the Endeavour Segment, Juan de Fuca Ridge: Correlations with seismicity and hydrothermal activity. *J. Geophys. Res.*, 112, B02401, doi: 10.1029/2005JB004210.
- Van Everdingen, D.A., 1995. Fracture characteristics of the sheeted dike complex, Troodos ophiolite, Cyprus: implications for permeability of oceanic crust. *J. Geophys. Res.* 100, 19,957-19,972.
- Wilcock, W.S.D., 2004. Physical response of mid-ocean ridge hydrothermal systems to local earthquakes. *Geochem. Geophys. Geosyst.* 5, Q11009, doi:10.1029/2004GC000701.
- Wilcock, W.S.D., Hooft, E.E.E., Toomey, D.R., McGill, R.R., Barclay, A.H., Stakes, D.S., Ramirez, T.M., 2009. The role of magma injection in localizing black-smoker activity. *Nature Geosci.* 2, 509-513, doi:10.1038/NGE0550.

Figure Captions

Figure 1. Phase diagram for NaCl-H₂O. (from Driesner et al., 2007).

Figure 2. Schematic of the geometry used in the numerical simulations showing applied boundary condition, crustal permeability, cell size, and scale dimensions.

Figure 3. (a) Temperature distribution with arrows showing fluid flow path after 50 years of simulation time for a system with constant basal heat flux of 130 W/m² in the left-most 500 m. Heat flux decreases linearly from 100 W/m² to 1 W/m² from 500 m to the right-hand edge of the system. (b) Green color depicts seawater salinity; blue color depicts plume salinity less than seawater; brown depicts salinity greater than seawater. The panel insert shows a magnified region of salinity associated with phase separation in the lower left-hand corner of the domain.

Figure 4. (a) Bottom temperature, (b) bottom salinity, (c) surface temperature, and (d) surface salinity as a function of time at selected distances from the left-hand boundary for the simulation described in Figure 3.

Figure 5. (a) Bottom temperature, (b) bottom salinity, (c) surface temperature, and (d) surface salinity as a function of time at selected distances from the left-hand boundary for a simulation with a sinusoidal heat flux of the form $100 + 50 \sin 2\pi t$ W/m².

Figure 6. (a) Bottom temperature, (b) bottom salinity, (c) surface temperature, and (d) surface salinity as a function of time at selected distances from the left-hand boundary for a simulation with a periodic box-shaped heat input at the base.

Figure 7. Heat flux profiles from a convecting, crystallizing magmatic sill of initial thickness of 100 m from Liu and Lowell (2009) both without replenishment and with replenishment at a rate

of 1.8×10^{-8} m/s. Because the initial heat fluxes are exceptionally high, we start the simulations after the heat flux has declined to the point indicated by the open circles on the curves.

Figure 8. (a) Bottom temperature, (b) bottom salinity, (c) surface temperature, and (d) surface salinity as a function of time at selected distances from the left-hand boundary for a simulation with a heat flux resulting from a convecting, crystallizing, unreplenished axial magma chamber

Figure 9. (a) Bottom temperature, (b) bottom salinity, (c) surface temperature, and (d) surface salinity as a function of time at selected distances from the left-hand boundary for a simulation with a heat flux resulting from a convecting, crystallizing, axial magma chamber that is replenished at a rate of 1.8×10^{-8} m/s.

Figure 10. Discrete measurements of (a) temperature and (b) salinity at Bio9, Bio9' and P vents at EPR 9°50' N between 1991 and 2008, spanning two eruptive events in 1991-1992 and 2005-2006. (from compilation of N. EPR vent fluid temperature, chloride, pH and Min. Mg values K.L. Von Damm, J.G. Bryce, and M.F. Prado, MGDS database)

Figure 11. Discrete measurements of (a) temperature and (b) salinity at the Main Endeavour Field between 1991 and 2005, including a non-eruptive diking event in 1999. Data from 1988 are from Butterfield (1994), 1990-2000 are from Lilley et al. (2003), and 2005 are from Foustoukos et al. (2009). (Reproduced from Singh et al., 2013)

812

813

814

815 **Figure 1**

816

Figure 2

825

826 (a)

(b)

827

828 **Figure 3.**

829

830

831

832

833

834

Figure 4

843

844

845

846 **Figure 5**

847

Figure 6

Figure 7

Figure 8

Figure 9

(a)

(b)

Figure 10

(a)

(b)

Figure 11

896 **Table 1.** List of parameters used in the equations, their definition, values, and units.

Symbol	Definition	Value	Units
a^*	Effective thermal diffusivity	10^{-6}	m^2/s
c	Specific heat	10^3	$\text{J/kg}\cdot^\circ\text{C}$
d	Depth of the system	1500	m
D	Diffusivity of salt	10^{-9}	m^2/s
g	Acceleration due to gravity	9.8	m/s^2
h	Specific enthalpy		J/kg
H	Heat flux		W/m^2
k	Permeability	10^{-13}	m^2
P	Pressure		Nt/m^2
S	Volume saturation		
t	Time		s
T	Temperature		$^\circ\text{C}$
v	Darcian velocity		m/s
X	Salinity		wt%
	<i>Greek Symbols</i>		

α	Thermal expansion coefficient	10^{-3}	$1/^{\circ}\text{C}$
ϕ	Porosity	0.1	
λ_m	Thermal conductivity of the porous medium	2	W/m.C°
μ	Dynamic viscosity of fluid		$\text{Pa}\cdot\text{s}$
ν	Kinematic viscosity of the fluid	10^{-7}	m^2/s
ρ	Bulk density of the fluid		kg/m^3
τ	Conductive time scale		s
ω	Angular frequency		s^{-1}
	<i>Subscripts</i>		
l	Liquid		
r	Rock		
v	Vapor		

897

898

899

900