ABSTRACT

We study one large submarine mud volcano that lies 210 km south-south east of New Orleans, Louisiana in 1070 meters of water depth. The vent is elevated approximately 75 meters relative to the surrounding seafloor. The vent is bounded by a negative polarity seismic reflection. We interpret that this reflection records the boundary from hydrate above to gas below: it is the Bottom Simulating Reflector (BSR). This BSR rises rapidly at the boundaries of the vent and is then horizontal within a few meters of the seafloor across the vent before it plunges once again on the far side. High temperature gradients and elevated salinities are present within the vent [1, 2]. We develop a 1D multiphase flow model to test whether the upward flow of fluids from depth can drive the coexistence of high salinity fluids, elevated temperature gradients, and an uplifted BSR. In this model, saline water advects heat and salt, whereas methane transports only heat. We show that salinity and temperature observations cannot be explained by the upward flow of saline water alone from a fixed depth. The observations can be explained, however, if methane is also advecting heat but not transporting salt and if the methane flux is 34 times larger than the water flux. Our results demonstrate that gas-dominated flow from depth may drive observations of elevated salinities and temperatures at vents in the GoM. A better understanding of the hydrogeological processes at vent zones is important for estimating fluxes of heat and mass and is important for understanding the conditions under which deep-sea biological communities exist.

Keywords: natural gas hydrates, seafloor vents, Northern Gulf of Mexico, salinity, temperature

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**NOMENCLATURE**

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<td>(\rho_w)</td>
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**Table 1.** List of parameters used in the models.

*\(T\)=temperature; \(M\)=mass; \(L\)=length; \(t\)=time.

### 1 INTRODUCTION

Gas hydrate is an ice-like compound that contains methane and other light gases in a lattice of water molecules [3]. It forms at high pressures and low temperatures and is ubiquitous on continental margins and in the Arctic where there are low temperatures and high pressures and sufficient supplies of gas and water. Elevated salinities and temperatures inhibit hydrate formation [3].

In the Northern Gulf of Mexico (GoM), hydrates have been observed in near-seafloor sediments at gas vent locations on the continental slope [4]. The Bottom Simulating Reflector (BSR), which marks the boundary between hydrate above and gas and water below, rises toward the seafloor at the center of the vent. Paull et al. [1] and Ruppel et al. [2] documented elevated temperatures and salinities near the seafloor of these vents. Although the presence of high salinities and temperatures and of gas hydrate have been well documented, the hydrogeological processes creating these conditions remain poorly understood.

Some researchers have proposed that high salinity values at vents result from either diffusive [5] or advective processes [6]. Others have shown that high salinities can be generated by the formation of authigenic gas hydrates [7] or by a subcritical phase separation [8]. Similar to explanations for high salinities, authors have applied an analytic solution to the steady-state advection-diffusion heat equation [9], suggesting that warm advecting fluids are elevating temperatures at vents on the seafloor. Others suggest that the conductive effects of an underlying salt body may help explain the observed elevated temperatures [9]. A separate study that uses a coupled model of fluid-heat-salt transport shows how salt-body geometry and seafloor relief may also spur salt-driven, geothermal convection on a regional scale [10]. Although several processes have been recognized that may contribute to increases in salinity and temperature gradients at seafloor vents, a dominant process that can explain both observations has yet to be fully described.

We characterize a vent in lease blocks MC852/853 in the Ursa Basin, where two studies have observed an uplifted BSR and measured subsurface salinity and temperature [1, 2]. We model the coexistence of high salinity fluids, elevated temperatures, and an uplifted BSR with two approaches:

1. We assume that warm, saline water is generated by dissolution of salt bodies at depth and that this hot, saline fluid is expelled vertically.
2. We model the multiphase expulsion of methane and saline water from depth. In this
model, water advects salt and heat, whereas methane advects only heat. This study is unique because we integrate both salinity and temperature data to test the transport-from-depth hypothesis. We present a model that explains observed salinities and temperatures and that constrains the flux of water and gas from vents.

2 VENT CHARACTERIZATION

At the seafloor, this vent is roughly circular with a diameter of 1.2 km. It projects above the surrounding sediment by ~75 m (figure 2). A seismic reflection profile across the vent shows that it is bounded by a negative polarity event (white) that is opposite to the polarity of the seafloor reflection (black) (figure 2). We interpret that this BSR records a negative impedance contrast that marks the boundary from hydrate and water above to gas below. What is striking is that this BSR rises rapidly at the boundaries of the vent and is then horizontal within a few meters of the seafloor before it plunges once again (figure 2).

The focus of our study is in the Ursa Basin located 210 km SSE of New Orleans (Figure 1). Ursa is a prolific hydrocarbon province that has been characterized by rapid sedimentation and regional fluid flow [11]. The region is spotted with topographic pockmarks, representing submarine mud volcanoes or seafloor vents that are actively releasing water and hydrocarbons (figure 2). These vents are the site of thermogenic gas hydrate deposits and biological communities [12, 13]. We focus on the most prominent vent of the region at lease blocks MC852/853 (figure 3).
Seward Johnson (IODP #554). These expeditions separately studied the vent at MC852/853 and documented significant thermal anomalies and high salinities [1, 2] (figure 3). Thermal gradients were measured in the upper ~3 m of sediment using the Dalhousie University heat flow probe equipped with 32 thermistors [2]. Porewater salinities were measured using a giant Calypso Piston Corer [1] and a Bentho 2450 Piston Corer [2]. Using box cores, Paull et al. [1] retrieved vein-filled, massive, structure II hydrate layers on the vent, indicating the presence of significant quantities of gas.

In the vent’s subsurface, salinities increase from seawater salinity at the seafloor to >4 times seawater salinity over just a few meters depth (figure 4). Away from the vent, measurements record dramatically different salinity-depth profiles. The profile from location PC-27, ~1.8 km from the vent’s center, shows no increase in salinity with depth (figure 4), whereas profiles from locations MD02-2565, MD02-2563, PC-25, and PC-26 each have a distinct concave-down shape (figure 4). Subsurface temperature measurements also show distinct variations. Thermal gradients increase from 24.5±0.8 mK m\(^{-1}\) to 435±43 mK m\(^{-1}\) over a 1.8 km transect approaching the vent (figure 5). Salinities and temperatures are significantly elevated at the vent’s core, but these anomalies are not recorded in measurements taken away from the vent.

Figure 3. Dip map of the mud volcano overlain with 25m depth contours at MC852/853. Image and contours are created using a 3-D exploration Seismic Survey. This dip map emphasizes steep gradients on surfaces such as faults and edifices. Steeper features are in black, and flatter features are colored gray. White diamonds correspond to piston core measurements and red circles to thermal measurements made by Ruppel et al. [2] and Paull et al. [1].

Figure 4. Salinities measured by piston cores. Data MD02-2563 and MD02-2565 were collected with a giant Calypso Piston Corer by Paull et al. [1], and piston core measurements PC25, PC26, and PC27 were collected by Ruppel et al. [14]. Salinity data are located on figure 3. PC25, MD02-2565, and MD-02563 were collected on the vent’s edifice. PC-26 was retrieved from the vent’s rim, and PC-27 was taken on the southern flank.

Figure 5. Figure adapted from Ruppel et al. [2] showing thermal gradients and associated errors measured from upper ~3 m of seafloor sediment [14]. Measurements are located in figure 3.

3 MODEL DESCRIPTION

We test the hypothesis that salinity and heat are advected from depth to create the observed profiles.

3.1 Salinity model. We combine conservation of mass for steady fluid flow with Fick’s law of chemical diffusion to generate the steady-state,
advection and diffusion equation for one-dimensional solute transport:
\[
0 = \frac{\partial^2 C}{\partial z^2} - \frac{V_z}{D} \frac{\partial C}{\partial z} \quad (1),
\]
where \( C \) is salinity, \( z \) is depth below seafloor, \( V_z \) is the Darcy velocity of saline water, \( \phi \) is porosity, and \( D \) is the molecular diffusion coefficient. We assume constant-salinity at the seafloor \( (C_u) \) and at the source depth \( (C_L) \). Ingebritsen et al. [15] presented the solution to equation 1 for constant-salinity boundary conditions:
\[
C(z) = C_u + (C_L - C_u) \frac{e^{\xi z/L} - 1}{[e^{\xi z} - 1]} \quad (2),
\]
where \( \xi \) is the dimensionless parameter:
\[
\xi = \frac{V_L \phi L}{D} \quad (3).
\]

3.2 Thermal model. In a similar fashion, we combine the conservation of mass and energy for steady fluid flow to generate the steady-state, advection-diffusion equation for one-dimensional heat transport:
\[
0 = \frac{\partial^2 T}{\partial z^2} - \frac{\rho_c c_w}{k_e} V_z \frac{\partial T}{\partial z} \quad (4),
\]
where \( T \) is temperature, \( z \) is depth below seafloor, \( \rho_c \) is density of seawater, \( c_w \) is the heat capacity of seawater, \( k_e \) is the thermal conductivity of the sediment column, and \( V_z \) is Darcy velocity. We assume a constant-temperature boundary condition at the seafloor and a constant heat flux \( (\Gamma_L) \) boundary condition at source depth \( (L) \). The solution to this form is defined by Bredehoeft and Papadopulos [16] as:
\[
T(z) = T_o + \frac{\Gamma_L L}{B} \frac{[e^{\xi z/L} - 1]}{[e^{\xi z} - 1]} \quad (5),
\]
where \( T_o \) is the temperature at the seafloor and \( B \) is the peclct number defined as:
\[
B = \frac{\rho_c c_w V_L}{k_e} \quad (6).
\]

3.3 Modeling MC852/853. Table 2 lists the values for the different boundary conditions and material parameters used in equations 2 and 5. We use these equations to assess the sensitivity of salinity and temperature profiles to different fluid flow velocities. We assume that the salinity at seafloor \( (C_u) \) is that of seawater \( (35 \text{ ppt}) \) and that the salinity at a depth \( L \) \( (C_L) \) is equal to \( 130 \text{ ppt} \). A value of \( 130 \text{ ppt} \) represents a typical value for reservoir salinities at depth [17] and is also approximately equal to the maximum salinity observed in the subsurface measurements [1] (figure 4). We assume that the seafloor temperature \( (T_o) \) is \( 4.8^\circ C \) based on direct measurements of bottom-water temperature [18]. The basal heat flux \( (\Gamma_L) \) is assumed to equal 24.5 \text{ mW m}^{-2} based on the thermal conductivity \( (1 \text{ W m}^{-1} \text{ k}^{-1}) \) and the background thermal gradient \( (24.5 \text{ mK m}^{-1}) \) [1, 2].

The material parameters are constrained as follows. Thermal conductivity \( (k_e) \) is assumed to be \( 1 \text{ W m}^{-1} \text{ k}^{-1} \) based on direct measurements [18]. Porosity \( (\phi) \) is approximated to be 0.45 for both models. We assume salinity diffusivity to be constant as \( 2.5 \times 10^{-10} \text{ m}^2 \text{s}^{-1} \). It is determined from the \( \text{Cl}^- \) diffusivity and is scaled for diffusion in a porous medium [19, 20]. We assume that salinity and heat are sourced from a depth \( (L) \) equal to 1000 meters below seafloor (mbsf). This is one of the most difficult parameters to determine: we do not have seismic data at this depth. We choose 1000 mbsf because it is the depth at which matured salt diapirs reach neutral buoyancy [21].

<table>
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<td>( C_L )</td>
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<td>observed [1]</td>
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<td>( T_o )</td>
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<td>observed [22]</td>
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<tr>
<td>( \Gamma_L )</td>
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<td>Derived from ( k_e ) and background thermal gradient (TM 5-1, 24.5mKm$^{-1}$)</td>
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<td>( \phi )</td>
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<td>( D )</td>
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<td>( \text{Cl}^- ) diffusivity and scaled for diffusion in porous medium [20]</td>
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<tr>
<td>( L )</td>
<td>1000 m</td>
<td>Depth for neutral buoyancy of matured salt bodies [21]</td>
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</table>

Table 2. Lists of the boundary conditions and material parameters used in the models.

4 MODEL RESULTS
4.1 Upward flow of saline water. We solve equation 2 for the velocity ($V_z$) necessary to produce a salinity profile similar to PC-25 (figure 6) which had the steepest salinity change with depth and recorded the highest observed value of salinity. We use the initial conditions and parameters presented in Table 2 and find an upward velocity of 1.5 cm yr$^{-1}$ provides a reasonable fit to the observed salinities for this site (figure 6). Similarly, we solve equation 5 for the velocity ($V_z$) necessary to produce the highest observed thermal gradient near the vent (TM 5-9, figure 7). We find an upward velocity of 4.9 cm yr$^{-1}$ produces a temperature profile similar to that observed (figure 7). Interestingly, the upward flow of water necessary to produce the observed temperature with these models is approximately 3 times that necessary to produce the observed salinity (4.9 cm yr$^{-1}$ vs. 1.5 cm yr$^{-1}$, figure 8).

**Figure 6.** An upward flow rate of 1.5 cm yr$^{-1}$ (red solid line) produces a salinity profile most similar to the observed salinity at PC-25 (blue solid line). Modeled salinity is based on equation 2 with parameters from table 2. 6a. Results plotted over a depth of 20 meters. The dotted black line indicates the depth extent expanded in 6b. 6b. Results plotted over a depth of 3 meters.

**Figure 7.** An upward flow rate of 4.9 cm yr$^{-1}$ (red solid line) produces a temperature profile most similar to the observed gradient at TM 5-9 (solid line with open blue dots). The interconnected open blue dots represent the thermal gradient (TM 5-9) extrapolated to a depth of 1000 m. This gradient is based on temperature measurements in the upper ~3 m of sediment (solid blue dots) and we do not believe it to persist at greater depths. Modeled
temperature is based on equation 5 with parameters from table 2. 7a. Results plotted over a depth of 1000 meters. The solid black box indicates the window dimensions in 7b. 7b. Results plotted over a depth of 50 meters.

Figure 8. An upward flow rate of 4.9 cm yr$^{-1}$ is necessary to produce the observed temperature profile whereas an upward flow rate of only 1.5 cm yr$^{-1}$ is necessary to produce the observed salinity. Cross-section B-B’ is located in figure 1b.

4.2 Model sensitivity analysis. We evaluate the effects of changing porosity and source depth on the estimated vertical velocities. Source depth and porosity are the two most poorly constrained parameters. They cannot be experimentally derived, and no field studies have determined these values for this vent location. Assuming 1000 mbsf source depth, we solve for vertical velocities with both the salinity and thermal model (equations 2, 5) for porosities ranging from 0.3 to 0.85. Similarly, assuming a 0.45 porosity, we estimate vertical velocities with the salinity and thermal model using source depths ranging from 1000 to 5000 mbsf.

As source depth (L) is increased from 1000 mbsf to 5000 mbsf, the flow velocity necessary to match the near-surface temperature gradient (figure 7b, TM 5-9) decreases from 4.9 cm yr$^{-1}$ to 0.98 cm yr$^{-1}$ (figure 9a). However, changing the source depth (L) for the salinity model does not significantly affect the flow velocity necessary to generate the near-surface salinity gradient (figure 6b, 9). Increasing porosity has an opposite effect on the thermal and salinity models, respectively. Increasing porosity causes the thermal model velocity to decrease and the salinity model to increase until the two predict equal velocities at 0.81 porosity (figure 9).

Our sensitivity analysis demonstrates that the observed salinity and temperature gradients could be explained by the upward flow of saline water from a source depth of 2990 mbsf (figure 9a). However, we find this unrealistic as this model results in an extraordinarily high temperature of 440°C at the source depth. In contrast, the regional temperature gradient of 24.5 mK m$^{-1}$ suggests a temperature of ~78°C at 2990 mbsf depth. Alternatively, the surface salinity and temperature gradients can be explained by the same flux of salty water from a depth of 1000 mbsf with a porosity value of 0.81. This value is
unreasonably high. In summary, these poorly constrained parameters significantly affect our estimates of water flow; however, for realistic source depths and porosities, the velocity required to create the observed temperature profile will be larger than the velocity required to replicate the observed salinity profile.

4.3 Upward flow of saline water and methane. As a preliminary attempt to explain the discrepancy in upward flow rates required to create the observed salinity and temperature profiles (figure 8), we hypothesize that methane is significantly contributing to heat anomalies at MC852/853. Specifically, we assume that thermal measurement TM 5-9 (435 mK m$^{-1}$) is created by the upward flow of gas and seawater. We also assume that the salinity gradient away from the vent at PC-25 is created by the upward flow of just brine.

Our assumptions are that: 1) gas and water are immiscible; and 2) heat transport is the sum of two independent transport pathways, through the gas and water phases. We combine conservation of energy and mass for steady flow to derive a multiphase steady-state, advection-diffusion equation for one-dimensional heat transport (appendix). This model is analogous to equation 4:

$$0 = \frac{\partial^2 T}{\partial z^2} - \left( \frac{q_g \rho_g c_g \phi}{k_e} + \frac{q_w \rho_w c_w \phi}{k_e} \right) \frac{\partial T}{\partial z}$$ (7).

In this model, $q_g$ and $q_w$ are the volume fluxes of seawater and gas, respectively. We assume a constant methane density, $\rho_g$ (85.5 kg m$^{-3}$), and heat capacity, $c_g$ (3242 J kg$^{-1}$ K$^{-1}$). The remainder of the variables are the same as for equation 4.

Using equation 7, we can calculate a range of solutions for gas and water flux that meet the surface temperature gradient (TM 5-9). If we assume the salinity model accurately predicts the upward flux of saline water ($q_w$) to be 1.5 cm yr$^{-1}$ (figure 6), then an additional gas flux ($q_g$) of 51 cm yr$^{-1}$ is required to create the observed 435 mK m$^{-1}$ thermal gradient. This calculation suggests that the methane flux is 34 times greater than the water flux.

This method is an initial and tentative attempt to explain the discrepancy in salinity and temperature gradients (figure 8). This model ignores well-understood behavior of multiphase flow such as: the inhibition of the flux of one phase due to the presence of another [15]. Future work, however, will focus on developing a coupled two-phase flow model that accounts for relative permeabilities of the gas and water phases.

5 SEISMIC AND MODEL AGREEMENT

By compiling salinity, temperature, and seismic data, we test the agreement of predicted and actual BSR depth at the center of MC852/853. We predict the BSR depth at the vent’s center by using salinity and gas composition measurements made at MC852/853 [1, 23]. Based on measured salinity gradients (e.g. M02-2565), we assume that porewater salinity remains constant at 13% at deeper depths. Using the Sloan (1998) CSMHyd Hydrate Model, we calculate hydrate equilibrium conditions for waters of 13% salinity assuming the gas is comprised of C1 (76.2%), C2 (9.0%), C3 (9.6%), i-C4 (4.7%), and n-C4 (.1%) [23]. We convert two-way travel time (ms) to depth (m) on the seismic reflection profile by assuming that the acoustic velocity is 1.484 km s$^{-1}$ above the seafloor (1070 mbsf) and that the acoustic velocity is 2 km s$^{-1}$ below the seafloor at measurement TM 5-9 (figure 10b). We relate depth (m) to pressure (MPa) by assuming that pressure increases hydrostatically through the sediment column (figure 10a). We assume temperature changes with depth as predicted by the thermal advection-diffusion equation matched to the 435 mK m$^{-1}$ thermal gradient (TM 5-9) (equation 5, figure 7). The intersection of the hydrate equilibrium curve with the modeled temperature data predicts the BSR depth (figure 10a). At the location of measurement TM 5-9, we calculate a BSR depth of 29 mbsf which is in agreement with the observed BSR (figure 10a,b).

In conclusion, there are temperature measurements to ~3 mbsf (figure 5). However, we estimate the BSR to be 29 mbsf (figure 10a). Our thermal model (equation 5) fits the shallow temperature data and predicts temperature at the deeper results. The interesting result from our thermal model is that it predicts that the BSR should lie 29 mbsf (figure 10b). This is indirect evidence to suggest that there is some validity to the model results.
Figure 10a. Plot showing thermal model results (red) (figure 7) and hydrate equilibrium curve (blue) for 13% salinity and for the specific gas composition at MC852/853. Red triangle represents the seafloor temperature and depth of thermal measurement TM -5-9. The intersection of the hydrate equilibrium curve with the modeled temperature/depth profile estimates the BSR depth to be 29 mbsf. 10b. Seismic reflection profile overlain with seafloor depth of measurement TM 5-9 (black) and predicted BSR depth (yellow) from figure 10a. This estimated depth (29 mbsf) in yellow coincides well with the depth of the BSR observed in seismic at the center of the vent.

6 DISCUSSION

Previous studies of seafloor vents have also estimated fluid fluxes based on salinity or temperature data. Ruppel et al. [2] study seafloor vents within the GoM, including the one we study here, and use a one-dimensional model to estimate upward flow rates of 15 cm yr\(^{-1}\) based on salinity measurements at PC-25 and 5 cm yr\(^{-1}\) based on salinity measurements at PC-26 (figure 4). Hornbach et al. [9] study a seep above the Blake Ridge Diapir off the U.S. east coast and apply an analytical solution to the steady-state advection-diffusion equation to estimate fluid flux based on observations of the BSR. They estimate fluid velocities through the vent ranging from 2 to 40 cm yr\(^{-1}\) which are ~100 times larger than previous estimates based on porewater geochemical data [9]. In both of these models, the flow of water is the only vehicle of heat or salt transport.

Our salinity and thermal models are similar to those from Ruppel et al. [2] and Hornbach et al. [9]. For salinity, we can replicate a velocity of ~15 cm yr\(^{-1}\) from Ruppel et al. [2] if we use a diffusion coefficient for Cl\(^-\) in water of 2.03x10\(^{-9}\) m\(^2\) s\(^{-1}\) [19]. However, for our models we prefer to use a diffusion coefficient for Cl\(^-\) scaled for a porous medium of 2.5x10\(^{-10}\) m\(^2\) s\(^{-1}\) [19, 20, 24], which yields a fluid velocity of 1.5 cm yr\(^{-1}\). In contrast to our estimated 1.5 cm yr\(^{-1}\) fluid velocity determined from salinity data, an upward velocity of 4.9 cm yr\(^{-1}\) is necessary to explain the observed temperature gradient at the seafloor (figure 8). The discrepancy between the two estimates is similar to previous studies that have determined that the flux constrained with temperature data is often greater than the flux rates constrained by porewater geochemical data [8, 9]. This result indicates that a model of simple advective flux of saline water cannot explain both temperature and salinity data.

Although our sensitivity analysis demonstrates that the upward flow of saline water from 2990 mbsf can create the observed salinity and temperature gradients (figure 9), this model requires an unrealistically 440\(^\circ\)C temperature at the source depth. In addition, we demonstrate that a 0.81 porosity will cause the salinity and thermal models to predict equal velocities. This porosity, however, is also highly improbable, especially given the 1000 mbsf source depth. This analysis is a further confirmation that there is a fundamental mismatch in the upward velocities necessary to explain observed temperature and salinity gradients.

We propose two possible explanations for the difference in flux estimates based on salinity versus temperature:

1) Another fluid besides saline water is being expelled from the vent. This fluid—likely methane—is advecting heat, but not transporting solute.

2) Separate processes are elevating salinity and temperature. For example, gas is advecting heat from a deep source, and salinity is generated via the exclusion of salt as hydrate forms [7, 25].

We explored the first hypothesis that methane and water are advecting heat while only
water is carrying salt. Our preliminary results suggest that the observed salinities and temperatures can be explained by the upward flow of both methane and saline water. In this model, saline water advects heat and salt, and methane advects only heat. We estimate that methane flux is ~34 times greater than the water flux.

We justify our multiphase flow model with four arguments: (1) studies have documented the expulsion of free gas at vents in the GoM [13] (2) the vertical wipe-out zone in the seismic reflection profile in the core indicates the presence of gas (3) the sulfate-methane transition (SMT) on the vent coincides with seafloor, suggesting a high vertical gas flux [1, 2] (4) the presence of high-saturation structure II hydrates at the vent implies that gas is being supplied in excess of the proportion of methane in hydrate [25].

It’s important to consider that the high gas flux at MC852/853 may indirectly drive salinity generation via the exclusion of salt as hydrate forms [7, 25]. While we assume that salinity is transported from depth by saline water, it is possible that salinity is elevated entirely as a result of hydrate formation. Future work will focus on developing this second hypothesis and comparing it to the results from this work.

We demonstrate that salinity and temperature anomalies at MC852/853 can be explained by the upward flow of gas and saline water from depth. Furthermore, our results and interpretation are significant because they reveal a method for determining gas and water flux from vents by measuring salinity and temperature. Our results indicate that methane flux is high relative to water flux.

7 CONCLUSION

Elevated salinities and temperatures have been observed at MC852/853. These salinities and temperatures appear to be confined to the area of the vent. We describe these phenomena with quantitative models of heat and salt expulsion from depth. We show that observed salinities and temperatures can be explained by upward flow of saline water from depth if in addition to the upward flow of water, a large flow of methane is also advecting heat but not transporting salt. We present a model that explains both observed salinities and temperatures at vents using the transport-from-depth hypothesis.

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APPENDIX

Derivation of equation 7. We assume that gas and water flux and the cross-sectional area of the vent conduit are constant. We define the water and gas fractions as $S_w$ and $S_g$, respectively. We assume heat transport is the sum of two independent transport pathways, through each of the two phases. From conservation of heat energy,

$$\Delta T \rho c_s \Delta z = \Delta q \Delta t$$  \hspace{1cm} (7.1).

For an advective-diffusive problem with two immiscible phases, we use conservation of mass and energy to define the total heat flux ($q$) as follows:

$$q = k_e \frac{\partial T}{\partial z} + (\nu_S \omega c_f + \nu_g \omega c_g) \nabla \phi$$  \hspace{1cm} (7.2).

Where the first term on the right side of the equation describes the heat diffusion due to the temperature gradient and the rest of the terms describe the advective heat flow through both water and gas. Substituting equation 7.2 into 7.1:

$$\frac{\Delta T}{\Delta t} = K_e \frac{\Delta \nabla T}{\Delta z} + (\nu_S \rho c_f S_c + \nu_g \rho c_g S_c) \frac{\partial \phi}{\partial z}$$  \hspace{1cm} (7.3),

where

$$K_e = \frac{k_e}{\rho c_s}$$  \hspace{1cm} (7.4),

$$k_f = \frac{\rho d c_f}{\rho c_s}$$  \hspace{1cm} (7.5),

and

$$k_g = \frac{\rho d c_g}{\rho c_s}$$  \hspace{1cm} (7.6).
\( K_s \) is thermal diffusivity; \( \rho_s \) is grain density; \( c_s \) is grain heat capacity. The remainder of the variables are defined in Table 1.

Because \( \Delta z = v \Delta t \) and \( v \) is a finite number, as \( \Delta t \to 0 \), we have \( \Delta z \to 0 \). We take \( \Delta t \to 0 \) on the left-hand side of the equation and \( \Delta z \to 0 \) on the right-hand side:

\[
\frac{\partial T}{\partial t} = K_e \frac{\partial^2 T}{\partial z^2} + (V_z S_o k_k + V_o S_y k_y) \frac{\partial T}{\partial z} \tag{7.7}.
\]

Substituting 7.4, 7.5, and 7.6 into 7.7 and considering only the steady-state situation where \( \partial T/\partial t = 0 \), we get:

\[
0 = \frac{\partial^2 T}{\partial z^2} - \left( \frac{V_z S_o \rho_o c_o \phi}{k_o} + \frac{V_o S_y \rho_y c_y \phi}{k_o} \right) \frac{\partial T}{\partial z} \tag{7.8}.
\]

By definition \( V_g S_g = q_g \) and \( V_w S_w = q_w \), where \( q_g \) and \( q_w \) are assumed to be constant. These fluxes are the volumetric flux over the entire area of the vent. Thus, we can simplify equation 7.8 to equation 7:

\[
0 = \frac{\partial^2 T}{\partial z^2} - \left( \frac{q_o \rho_o c_o \phi}{k_o} + \frac{q_w \rho_w c_w \phi}{k_o} \right) \frac{\partial T}{\partial z} \tag{7}.
\]