Introduction

The following supplementary material contains an extended discussion of the multiphase evolution of the roofs and floors of sills simulated and discussed within the main manuscript and discusses the influence of reservoir chemistry and thermal driving on the thicknesses and propagation rates of these interfaces (Text S1). Additionally, we include a sensitivity study (Text S2), which explores the impact of variable material properties (thermal conductivity and dynamic viscosity) and spatially heterogeneous thermal boundary conditions (linear temperature profiles along fracture sidewalls) on the resultant bulk salinity profiles of resolidified sills and fractures. Figure S1 depicts the temporal evolution (growth) of sill floors and roofs during sill solidification. Figure S2 depicts the chemical evolution (salination) of the residual liquid reservoir during sill solidification. Figure S3 depicts variations
in sill (Figure S3a) and fracture (Figure S3b) bulk salinity profiles caused by amplified ice thermal conductivity and dynamic viscosity. Figure S4 depicts variations in fracture bulk salinity profiles when linearly varying sidewall temperatures are used to drive solidification (as opposed to the isothermal boundary conditions implemented in the main text). Captions for Movies M1-M33 describe the simulations these supporting .avi movies depict.
Text S1 – Sill Roof and Floor Evolution

During solidification the physicochemical properties and propagation rates of the multiphase ‘mushy layers’ that characterize the ice-brine interfaces of the roofs and floors of sills evolve. This evolution is dependent on the thermal gradients driving the solidification as well as the chemical composition of the sill. In Figure S1 we show the temporal evolution (growth) of the roof and floor mushy layers of the two sills (35 ppt NaCl and 35 ppt MgSO₄) we simulated in the main manuscript as well as the roof and floor mushy layer evolution of two additional sill simulations (1 – a freshwater (0 ppt) sill subject to the same thermal driving as the sills in the main manuscript and 2 – a 35 ppt NaCl sill with symmetric undercooling of 132 K at both the top and bottom boundary). These simulations were carried out to investigate the impacts environmental conditions (e.g., thermal driving, initial reservoir chemistry) have on mushy layer and sill roof/floor evolution. In three of the simulations, we track the ice-mush (IM) phase boundary as well as the mush-liquid (ML) phase boundary for both the roofs and floors of the sills. The space between these two phase boundaries defines the mushy layer – where a nonzero brine volume fraction exists. There does not exist a mushy layer in the freshwater sill (an expected result of freezing a pure fluid [Huber et al., 2008; Rubinšteĭn, 2000]) so only the ML phase boundary is shown and represents a sharp ice-water transition.

Several expected trends are apparent in Figure S1 including 1) larger undercoolings lead to faster interface propagation; 2) floors have thicker mushy layers than roofs (consistent with our conclusions in the main manuscript); 3) mushy layers thin near the end of the sill solidifications as the residual fluid concentrates; and 4) mushy layers in MgSO₄ systems are much thinner than those of NaCl systems. Another less intuitive trend is also evident – the similar propagation rate of the freshwater sill roof and the ML interfaces of comparably undercooled saline sills, which sometimes even exceed the rate of the freshwater sill. While somewhat counterintuitive, given the freezing point depression effects of a saline/concentrating sill, ML interface propagation is primarily driven by conductive heat loss to the cold adjacent ice [Buffo et al., 2021a], which can continue to be efficient in the ice phase of the mushy layer. Brine convection within the mushy layer may also amplify the efficiency of heat loss from the liquid reservoir, potentially explaining the ML propagation rates that exceed those of the freshwater roof. Interestingly, the ML interface propagation of the sill floor outpaces that of the sill roof in the symmetric undercooling case (dashed blue line and solid blue line of Figure S1, respectively). This suggests that while brine convection in the roof mushy layer amplifies heat loss from the residual reservoir it also acts to cycle relatively warm water into the roof mushy layer from the reservoir, slowing its ML interface propagation in relation to the floor ML interface.
In the end these comparative simulations show that both thermal driving and reservoir chemistry play a role in governing mushy layer thicknesses – shallower thermal gradients lead to thicker mushy layers and mushy layer thickness is proportional to the freezing point depression effects of solutes (i.e., NaCl results in thicker mushy layers as it has a much low eutectic temperature and saturation point). Conversely, while thermal driving plays a large role in governing mushy layer interface propagation rates (interface propagation rates are proportional to the magnitude of the driving thermal gradient) salinity plays a much smaller role and minimally impacts the rate of the ice-liquid/mush-liquid interface (additional tests with nonzero salinities distinct from the 35 ppt values used in the current simulations would need to be carried out to determine the effect of salinity on the propagation rates of the IM interface).

Environmental conditions (e.g., thermal environment, brine chemistry) have significant impacts on the structure, dynamics, and evolution of ice-brine interfaces on icy worlds throughout the solar system, including Earth [Feltham et al., 2006; Hunke et al., 2011]. Given the importance of these interfaces in governing the evolution of planetary ice shells and any internal hydrological features they may contain as well as their integral role in mediating material and heat transport between planetary hydrospheres and cryospheres [Buffo et al., 2020; Vance et al., 2016; Vance et al., 2020], constraining the physicochemical properties and dynamics of ice-brine mushy layers will play a fundamental role in improving our understanding and predictive modeling capabilities of ice-ocean world geophysics, habitability, and spacecraft mission observations (see discussions in Buffo et al. [2021b] and Vance et al. [2020]).

Text S2 – Sensitivity Studies

Given the novelty of our simulations, we have implemented several model simplifications to identify and isolate first order patterns and dependencies in resultant ice properties caused by unique freezing front geometries. These simplifications include isothermal boundary conditions at the sidewalls of basal fractures, homogeneous and constant material properties (e.g., thermal conductivity, specific heat, ice density, molecular diffusivity, brine viscosity), use of the Boussinesq approximation, and solving Darcy’s law in the entire simulation domain (see [Buffo et al., 2021b] for a detailed discussion of the implications of such approximations and assumptions). These same simplifications are routinely employed by contemporary state-of-the-art sea ice models (e.g., [Parkinson et al., 2020; Wells et al., 2019]). However, in reality, the ice-ocean and ice-brine interfaces of hydrological features within planetary ice shells will likely exist as complex thermal, chemical, and physical environments with heterogeneous material
properties and complex/variable boundary conditions [Buffo et al., 2020; Buffo et al., 2021b; Chivers et al., 2021; Vance et al., 2020].

Here we explore the sensitivity of our model results to three variations in model boundary conditions and simulated material properties: 1) variation of ice thermal conductivity, 2) variation of brine viscosity, and 3) spatially variable thermal boundary conditions.

In planetary environments, low temperatures can influence the thermal conductivity of ice. Notably, thermal conductivity increases with decreasing ice temperature. Here we investigate the effects of increasing the thermal conductivity of ice to 4.0 W m\(^{-1}\) K\(^{-1}\) (representative of ice near 140 K [Wolfenbarger et al., 2021]), from its initial 2.0 W m\(^{-1}\) K\(^{-1}\), on resultant ice bulk composition (Figure S3). This is an extreme endmember case as thermal conductivity would vary gradually with ice temperature and remain at \(\sim\)2.0 W m\(^{-1}\) K\(^{-1}\) near any ice-ocean/brine interfaces where freezing and entrainment occurs. This was previously demonstrated by [Buffo et al., 2020], who showed that temperature dependent ice thermal conductivity has a negligible effect on ice-ocean interface thermal gradients and entrainment dynamics. This minimal variation in salt entrainment is caused by the interfacial mushy layer acting as a thermal and physicochemical buffer between the considerably different material property regimes of the completely solidified (\(T<T_{\text{eutectic}}\)) upper ice shell and completely fluid regions (ocean, liquid reservoirs – \(T>T_{\text{melt}}\)). Acting in a fashion akin to a low permeability strata layer restricting vertical liquid transport in the Earth’s crust, this layer’s properties can set a ‘speed limit’ for material and energy transport, even if other portions of the ice shell are capable of higher transport rates.

We use this endmember case study to demonstrate that even under artificially enhanced thermal conductivities ice bulk composition is minimally altered. We conduct similar sensitivity studies using an amplified dynamic fluid viscosity, as high salt concentrations exhibited by brines within the interfacial mushy layer could significantly increase their viscosities [Laliberte, 2007]. Here we set \(\mu=5.64\times10^{-3}\) Pa s, three times that of our baseline dynamic viscosity (Table 1) and in line with a nearly saturated magnesium sulfate solution near its freezing point (extrapolated from [Korosi and Fabuss, 1968]). Like our thermal conductivity study, this is an artificially high upper endmember case, and bulk salinity variations in actual ice-ocean world systems are expected to be less than those simulated here.

Horizontally averaged bulk salinity profiles of refrozen sills (initially filled with a 35 ppt NaCl solution) with amplified ice thermal conductivity and amplified dynamic viscosity are compared to that of the baseline case presented in the main text (Figure S3a). In both cases, the overall bulk salinity profile of the resolidified sill is minimally affected. The overall shape of the profile is consistent between all
simulations and the bulk salinity values are only slightly amplified in the two sensitivity studies. This is a logical result as an amplified ice thermal conductivity will lead to faster freezing of the sill (4,216 years versus the baseline 8,431 years; entraining more salt) and an amplified dynamic viscosity will reduce the efficiency of brine drainage into the residual brine reservoir, also entraining more salt (freeze out time of 8,431 years, the same as the baseline case). We additionally plot the absolute differences in bulk salinity values between our baseline results and the two sensitivity studies. The mean difference between the amplified thermal conductivity study and the baseline values is 1.87 ppt corresponding to a 12.9% difference and the median difference is 0.92 ppt corresponding to a 5.12% difference. The mean difference between the amplified dynamic viscosity study and the baseline values is 1.77 ppt corresponding to a 12.1% difference and the median difference is 0.95 ppt corresponding to a 4.71% difference.

We conducted similar sensitivity studies for two scales of 35 ppt MgSO$_4$ ocean-filled fractures – 10 m wide fractures and 500 m wide fractures. The horizontal bulk salinity profiles in the central regions of these two fractures (4.8 m from the top of the 10 m wide fracture and 48 m from the top of the 500 m wide fracture) and their comparison to the unmodified case of the main text can be seen in Figure S3b. While the salinity variations experienced by resolidifying fractures with amplified thermal conductivity and viscosity are larger than those experienced by the lenses (likely due to the increased residence time of cryoconcentrated brine in the vertical sidewall mushy layer as it percolates downward toward the ocean – See Section 4.2 of the main text), these variations, as endmember maximums, remain moderate. Mean variations in ice bulk salinity in the top half of these fractures range from 26.1-37.8% and median variations range from 25.7-38.3%.

These are the maximum possible variations in bulk ice salinity that can be caused by such heterogeneous material properties (thermal conductivity and dynamic viscosity) and the actual variations will likely be much lower than this. For example, minimum temperatures in the mushy layer will be set by the eutectic temperature of the system (~250 K for NaCl dominated brines and ~269 K for MgSO$_4$ dominated brines), thus even for the more heavily impacted NaCl system we would expect thermal conductivity and bulk concentration values in the interfacial mushy layer to vary by a factor of six less than observed in our sensitivity analysis (~1-6% mean difference across both lenses and fractures). This is further supported by the nonlinearity of variations in ice thermal conductivity, which varies less near the freezing point and more at colder temperatures [Potter et al., 2019]. Similar nonlinearities characterize the relationship between brine viscosity and salinity [Korosi and Fabuss, 1968; Laliberte, 2007], with viscosity increasing exponentially as solution concentration increases. The majority of the hydraulically connected brine within the mushy layers of hydrological features is not near its
eutectic concentration (see the bulk concentrations of brines within the brine channels and emanating plumes of mushy layers in Figures 5 & 6) and NaCl brines exhibit less amplification in viscosity than do MgSO₄ brines [Korosi and Fabuss, 1968]. As such, we would expect actual viscosity and bulk salinity variations to be at least 3 times lower than those investigated by our sensitivity studies (~2-12% mean difference).

This is not to say that such heterogenous material properties are unimportant. Quite conversely, as we begin to constrain ocean world properties with upcoming mission observations (e.g., ice compositions, ocean and brine compositions/concentrations, ice shell thermophysical structure) such nuances will be essential for interpreting the icy fingerprints left by geophysical and geological ice-brine processes and constraining the relationships between their remotely observable characteristics and the properties of subsurface fluid reservoirs [Howell and Pappalardo, 2018; Howell and Pappalardo, 2020; Schmidt, 2020; Schmidt and Buffo, 2017; Vance et al., 2020]. Rather, we emphasize that given the current uncertainties in first order ocean, brine, and thermal characteristics of ice-ocean worlds (e.g., ocean and brine composition/concentration, ice shell thermophysical structure) [Hand et al., 2007; Hand and Chyba, 2007; Zolotov, 2007; Zolotov and Shock, 2001] signatures of relatively dramatic variations in heterogeneous material properties could easily be lost amongst, for example, order of magnitude uncertainties in ocean concentration [Hand and Chyba, 2007]. Here we have demonstrated that variations in thermal conductivity and brine viscosity have a minimal impact on first-order structure and quantity of salt entrainment in planetary ice-brine systems when compared to other environmental unknowns such as ocean concentration (e.g., see the significant entrainment rate variations in cryoconcentrating sills – Figure 4). Future studies investigating the magnitude and uniquity of the signatures that these, and other, heterogeneous material properties produce will continue to increase the fidelity of such forward models and will play a fundamental role in the interpretation of upcoming spacecraft data interpretation (e.g., Europa Clipper, Dragonfly), particularly as first order uncertainties such as ocean composition and ocean concentration are further constrained.

In the main text we assume that basal fractures propagate into an isothermal region of an ice shell (e.g., the proposed ductile convective portion of Europa’s ice shell [McKinnon, 1999]). Given the novelty of our study, this was done to isolate the effects horizontal freezing front propagation has on resultant salt entrainment rates. If, however, such fractures were to propagate into conductive portions of an ice shell (e.g., above lenses into a brittle lid, into shells that are too thin to support solid state convection) their sidewall temperatures would be characterized by an approximately linear thermal profile (low temperatures near the fracture tip increasing to the ocean/reservoir temperature at the base of the fracture) [Buffo et
Here we briefly explore the impact of such a linear sidewall temperature profile on the resultant salt entrainment dynamics in basal fractures. Figure S4 depicts the bulk salinity profiles produced when a linear thermal profile is implemented along the fracture walls of a 10 m wide by 20 m tall MgSO$_4$ ocean-filled fracture (Figure S4a) and a 500 m wide by 1000 m tall MgSO$_4$ ocean-filled fracture (Figure S4b). In each case, the sidewall temperature is held constant throughout the duration of the simulation (Dirichlet boundary condition) and varies linearly from 200 K at the top of the fracture to 273.35 K (ocean temperature) at the domain’s vertical midpoint. Vertically averaged horizontal bulk salinity profiles for descending regions within the fractures are plotted (leftmost plots of Figure S4) to assess whether amplified salt retention with increasing depth is observed in the linearly varying sidewall temperature scenario, as it was in the isothermal baseline simulations.

Bulk salinities in the 10 m wide fracture initially increase with depth under linearly varying sidewall temperatures (like the baseline case), but then begin to decrease. However, bulk salinities in lower portions of the resolidified fracture remain above or near those of the uppermost region of the fracture. This is significant as it supports an important hypothesis of the main text, which suggests that increased residence time and cryocencentration of the brine in the interfacial mushy layer caused by a horizontally propagating freezing front geometry leads to amplified salt entrainment. If this were not the case, deeper regions of the fracture subject to much lower thermal gradients would be expected to entrain significantly less salt than upper portions of the fracture. In the 500 m wide fracture bulk salinities nearly always decrease with increasing depth (the exception is the very edges of the fracture, where larger thermal gradients are present) when linearly varying sidewall temperature are implemented. These results suggest that as thermal gradients decrease, and brine drainage becomes more efficient, the amplified residence time of the downwelling brine in the sidewall mushy layer has a dwindling impact on amplified cryoconcentration and resultant salt entrainment. This decrease in salt entrainment is additionally exacerbated by the tendency for the ice-ocean interface of fractures with linearly varying sidewall temperatures to deviate from a truly vertical wall geometry (e.g., the rightmost plot of Figure S4b). As the interface tends towards a more sloped geometry downwelling brine plumes are more likely to exit the mushy layer into the central fluid filled portion of the fracture before reaching the fracture base, reducing brine residence times in the mushy layer and thus salt entrainment. Similar trends in reduced material entrainment at sloped ice-brine interfaces have been observed experimentally [Leitch, 1987; Leitch, 1985].
As the geometric, thermal, physical, and chemical environmental conditions of ice-ocean and ice-brine systems become more complex so do their resulting entrainment trends. Variations in material properties, thermal driving, interface geometry, and freezing rate can have compounding, competing, and confounding effects on the ultimate material entrainment rates and resultant characteristics of forming ice. Isolating the impacts of each, understanding their prevalence across diverse ice-ocean worlds, and constraining the role they play in the geophysics, ice shell material transport capabilities, and observable properties of planetary ices will play a fundamental role in the upcoming exploration of high priority solar system bodies (e.g., Europa, Titan, Enceladus). Here we have begun to highlight how such variations can alter the rates of material entrainment in ice-brine systems and the magnitudes of such entrainment variations. Combined with our results from the main text regarding the important role of interface geometry, we have demonstrated that intrashell hydrologic features (e.g., lenses, fractures, dikes, sills) provide unique and dynamic environmental conditions capable of facilitating heterogeneous and amplified material entrainment in planetary ices that cannot be explained by contemporary models of material entrainment at a planar ice-ocean interface (e.g., [Buffo et al., 2020; Buffo et al., 2021b]).
Figure S1. The growth of sill floor and roof mushy layers. The temporal propagation of the key interfaces that define the mushy layers of sill floors and roofs are plotted for all four simulations described in Text S1. Lines labeled as ‘MgSO₄’ and ‘NaCl’ represent results from the 35 ppt sill simulations described in the main manuscript. Lines labeled as ‘Cold Base’ represent results from a 35 ppt NaCl sill solidification simulation driven by symmetric Dirichlet thermal forcing at its upper and lower boundaries of 132 K. Lines labeled as ‘0 ppt’ represent results from the solidification of a freshwater sill subject to the same thermal forcing describe in the main manuscript. ‘IM’ signifies the ice-mush interface – the transition between a solid below the eutectic (porosity = 0) and the mushy layer (porosity >0), and ‘ML’ signifies the mush-liquid interface – the transition between the mushy layer and the reservoir fluid (porosity = 1).
Figure S2. Salination of solidifying sills. As isolated sills freeze brine is rejected from the mushy layers of their roofs, concentrating their residual liquid reservoir. The temporal evolution of this process is shown for three different simulations. The plateaus near the end of the run correspond to the eutectic concentrations of the respective sills.
Figure S3. Material property effects on salt entrainment and bulk salinity. a) Variations in bulk salinity profiles of a resolidified NaCl ocean-filled sill when thermal conductivity and brine viscosity are amplified by a factor of two (red line) and three (yellow line), respectively. Absolute differences between the baseline control case (blue line - the scenario explored in the main text) and the sensitivity studies are also plotted (black lines). Minimal qualitative and quantitative variations are exhibited. (H = 1000 m) b) Similar sensitivity study results for MgSO₄ ocean-filled fractures (left – 10 m wide fracture [H = 10 m], right – 500 m wide fracture [H = 1000 m]). Lines represent horizontal bulk salinity profiles taken 4.8 m and 48 m from the tops of the fractures, respectively. The central portions of the fractures (slowest freezing rates) exhibit the largest discrepancies between the control simulation and our sensitivity studies.
Figure S4. Bulk salinity profiles in fractures with linearly varying sidewall temperatures. a) In a 10 m wide fracture (H = 10 m), horizontal bulk salinity profiles (left plot) are vertically averaged over the regions depicted in the two right plots (similar to Figure 8 of the main text, regions are bounded by their associated lines). These 2D profiles represent the resolidified fracture bulk salinity distribution when subject to constant 200 K sidewall temperatures (control scenario – center plot) and when subject to a linearly varying sidewall temperature (200 K at the top of the fracture [z/H = 4.0] and ocean temperature [273.25 K for MgSO₄] at the ice-ocean interface [z/H = 2.0] – right plot). In the linear sidewall temperature scenario, increasing bulk salinities with increasing depth are not nearly as prevalent as they are in the isothermal sidewall scenario, however there
are portions of the 10 m fracture where this is still the case. Additionally, even with significantly smaller thermal gradients driving solidification in lower parts of the fracture a nearly constant level of salt is entrained at all depths. b) Results from the same simulation approach as in panel (a) but for a 500 m wide fracture (H = 1000 m). In this wider fracture, except for immediately next to the fracture sidewall, bulk salinities always decrease with depth when a linearly varying sidewall temperature is employed.
**Movie S1.** Bulk salinity evolution of a 1 km thick 35 ppt NaCl sill subject to the undercooling boundary conditions presented in Figure 2. Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S2.** Porosity evolution of a 1 km thick 35 ppt NaCl sill subject to the undercooling boundary conditions presented in Figure 2.

**Movie S3.** Streamline evolution of a 1 km thick 35 ppt NaCl sill subject to the undercooling boundary conditions presented in Figure 2. Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S4.** Bulk salinity evolution of a 1 km thick 35 ppt MgSO$_4$ sill subject to the undercooling boundary conditions presented in Figure 2. Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S5.** Porosity evolution of a 1 km thick 35 ppt MgSO$_4$ sill subject to the undercooling boundary conditions presented in Figure 2.

**Movie S6.** Streamline evolution of a 1 km thick 35 ppt MgSO$_4$ sill subject to the undercooling boundary conditions presented in Figure 2. Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S7.** Bulk salinity evolution of a 1 m by 1 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S8.** Porosity evolution of a 1 m by 1 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

**Movie S9.** Streamline evolution of a 1 m by 1 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S10.** Bulk salinity evolution of a 10 m by 20 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S11.** Porosity evolution of a 10 m by 20 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

**Movie S12.** Streamline evolution of a 10 m by 20 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).
undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

Movie S13. Bulk salinity evolution of a 100 m by 200 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

Movie S14. Porosity evolution of a 100 m by 200 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

Movie S15. Streamline evolution of a 100 m by 200 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

Movie S16. Bulk salinity evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

Movie S17. Porosity evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

Movie S18. Streamline evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

Movie S19. Bulk salinity evolution of a 1 m by 1 m 35 ppt MgSO₄ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

Movie S20. Porosity evolution of a 1 m by 1 m 35 ppt MgSO₄ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

Movie S21. Streamline evolution of a 1 m by 1 m 35 ppt MgSO₄ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

Movie S22. Bulk salinity evolution of a 10 m by 20 m 35 ppt MgSO₄ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).
undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S23.** Porosity evolution of a 10 m by 20 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

**Movie S24.** Streamline evolution of a 10 m by 20 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S25.** Bulk salinity evolution of a 100 m by 200 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S26.** Porosity evolution of a 100 m by 200 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

**Movie S27.** Streamline evolution of a 100 m by 200 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S28.** Bulk salinity evolution of a 500 m by 1000 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S29.** Porosity evolution of a 500 m by 1000 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling).

**Movie S30.** Streamline evolution of a 500 m by 1000 m 35 ppt MgSO$_4$ ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (200 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.

**Movie S31.** Bulk salinity evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (260 K undercooling).
undercooling). Black contours demarcate porosities ranging from 0.15 to 0.95 in increments of 0.2.

**Movie S32.** Porosity evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (260 K undercooling).

**Movie S33.** Streamline evolution of a 500 m by 1000 m 35 ppt NaCl ocean filled fracture subject to the undercooling boundary conditions presented in Figure 2 (260 K undercooling). Streamlines are represented as blue to red contours that indicate relative flow speed along the streamline.
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