MODELING OF POST-IMPACT MELT INFILTRATION AND FREEZING ON GALILEAN SATELLITES . D. G. Korycansky, *CODEP, Department of Earth and Planetary Sciences, University of California, Santa Cruz CA 95064* .

Introduction

The Galilean satellites with icy surfaces (Ganymede, Callisto, Europa) are host to a variety of large impact features that are, if not unique to these bodies, rarely encountered on planetary and satellite surfaces in the Solar System. These features include impact basins with central pits, domes, and so-called “penepalimpsests” and “palimpsests” in the terminology of Schenk *et al.* 2004. It is likely that the unique morphology of these features is due to long term processes enabled by the material properties of the target surfaces.

One potential determinant of crater morphology is the possible infiltration of post-impact melt below the craters. Based on observations of terrestrial impact structures, lunar gravity data, and Martian crater morphology, we may expect extensive fractured zones to form underneath impact sites on Galilean satellites. Melt generated during the impact would be expected to percolate downwards into the zone beneath the target. This drainage may account for observed features such as central pits in craters. Models for infiltration and drainage of impact melt might explain observed features. For example, Elder *et al.* 2012 proposed an idealized model of cm-scale vertical crevasses that provided drainage of melt.

Infiltration modeling

An idealized starting point is to treat the target substrate as a variably saturated medium into which melt drains by infiltration. One widespread strategy that we borrow from the hydrology literature is a formulation in terms of the so-called Richards equation (e.g. Klute 1952, or recent discussions by Farthing and Ogden 2017, Zha *et al.* 2019):

$$\frac{\partial \theta}{\partial t} = \nabla \cdot [K(h)\nabla h] - \frac{\partial K}{\partial z}, \qquad (1)$$

where θ is the fluid concentration, $h = \Psi/g\rho < 0$ is the moisture potential scaled in terms of the so-called pressure head h, and $K(h)$ is the nonlinear conductivity for unsaturated porous flow. The second term on the right-hand side accounts for vertical gravity-driven drainage, parameterized by the conductivity K. We choose a simple power-law parameterization of $K(h) = K_s(h/h_s)^{\mu}$ for $h < h_s < 0$, where K_s is a nominal conductivity and $h_s < 0$ is the pressure head corresponding to saturation $\theta = \theta_0$. Given a relation between $\theta = \theta(h)$ between the concentration and the pressure head, the time-dependent term $\partial\theta/\partial t$ is commonly expressed in terms of h, i.e. $\partial\theta/\partial t = C(h)\partial h/\partial t$, where $C(h) = d\theta/dh$.

The Richards equation is a highly nonlinear equation of mixed type, parabolic in the unsaturated regime, and elliptic for saturated regions where θ equals the saturated value θ_s and $C(h) = 0$. Solution of the Richards equation is subject to some complications but is relatively straightforward if discretized

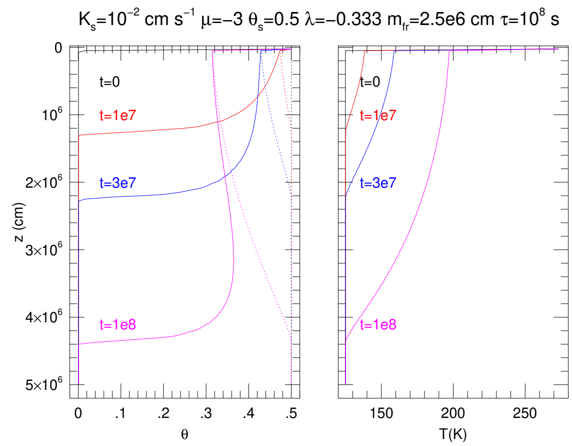

Figure 1: Solutions of eqns 8 for infiltration and freezing for a infiltration-dominated calculation ($Fz = 5$). The left panel shows the fluid fraction θ (solid curves) and porosity θ_0 (dotted curves) at various times. The right panel shows the ice temperature T_I at the same stages of the calculation.

with an implicit time-stepping scheme (cf. Sadegh Zadeh 2011).

Freezing can be incorporated using a simple model that we base on Illangasekare *et al.* 1990, where a heat transfer term Q mediates heat flow from melt to the surrounding ice substrate. We write $Q = C_I(T_m - T_I)/\tau$, where T_m is the melt temperature of water ice (273.15 K), T_I is the local temperature of the ice substrate, C_I is the heat capacity of the ice, and τ is a freezing timescale, treated as a parameter. The sink term to be incorporated in the Richards equation for freezing of the melt fraction θ is

$$\left.\frac{\partial \theta}{\partial t}\right)_f = -\frac{\theta Q}{L_m}, \qquad (2)$$

where L_m is the latent heat per unit mass of melting/freezing for water. The corresponding rate of change of the porosity in the ice is

$$\frac{\partial \theta_0}{\partial t} = \frac{\rho_m}{\rho_I}\left.\frac{\partial \theta}{\partial t}\right)_f, \qquad (3)$$

where ρ_m and ρ_I are the densities of melt and ice, respectively. Adding the source term in the heat-diffusion equation for ice temperature T_I we have

$$\frac{\partial T_I}{\partial t} = \frac{\partial}{\partial z}\left(\kappa_I \frac{\partial T}{\partial z}\right) + \frac{\theta(T_{melt} - T_I)}{\tau(1-\theta_0)}\frac{\rho_m}{\rho_I}, \qquad (4)$$

where κ_I is the heat conductivity of ice.

As θ_0 is now a function of time, an extra term appears in the relation of θ and h in the time-dependent term of the

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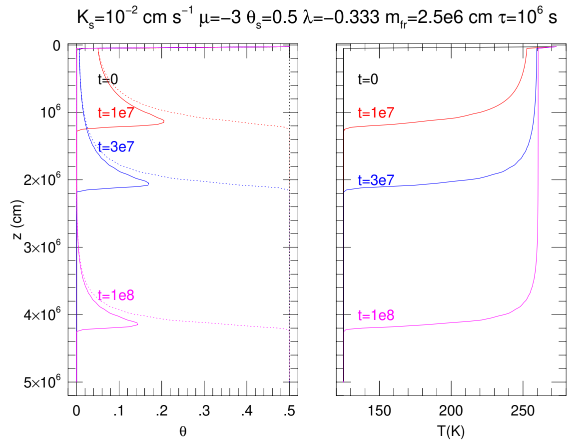

Figure 2: Solutions of eqns 8 for infiltration and freezing for a freeze-dominated calculation ($Fz = 500$). Left and right panels as in Fig. 1.

Richards equation. We assume a simple power-law relation for θ as a function of h: $\theta = \theta_0 (h/h_s)^\lambda$, giving us

$$\frac{\partial \theta}{\partial t} = \frac{\partial \theta_0}{\partial t}\left(\frac{h}{h_s}\right)^{\lambda} + \frac{\theta_0}{h_s}\left(\frac{h}{h_s}\right)^{\lambda-1}\frac{\partial h}{\partial t} = \frac{\rho_m}{\rho_I}\left.\frac{\partial \theta}{\partial t}\right)_f \left(\frac{h}{h_s}\right)^{\lambda} + C\frac{\partial h}{\partial t}. \tag{5}$$

Putting all the terms together in the Richards equation (where we specialize to the one-dimensional case in vertical direction), we have

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z}\left[K\left(\frac{\partial h}{\partial z} - 1\right)\right] + \left.\frac{\partial \theta}{\partial t}\right)_f \tag{6}$$

or

$$C\frac{\partial h}{\partial t} = \frac{\partial}{\partial z}\left[K\left(\frac{\partial h}{\partial z} - 1\right)\right] + \left.\frac{\partial \theta}{\partial t}\right)_f \left[1 - \frac{\rho_m}{\rho_I}\left(\frac{h}{h_s}\right)^{\lambda}\right]. \tag{7}$$

Equations 3, 4, and 7 form a system for the pressure head h, the porosity θ_0 and ice temperature T_I.

Nondimensionalized equations are

$$\begin{aligned}
\frac{\partial \theta}{\partial t} = C\frac{\partial h}{\partial t} &= \frac{\partial}{\partial z}\left[K\left(\frac{\partial h}{\partial z} - 1\right)\right] - \frac{S\theta(1-T)}{\tau}\left[1 - \delta\left(\frac{h}{h_s}\right)^{\lambda}\right], \\
\frac{\partial T}{\partial t} &= \frac{\partial}{\partial z}\left(\kappa_I \frac{\partial T}{\partial z}\right) + \frac{\delta}{(1-\theta_0)}\frac{\theta(1-T)}{\tau}, \\
\frac{\partial \theta_0}{\partial t} &= -S\delta\frac{\theta(1-T)}{\tau}.
\end{aligned} \tag{8}$$

The equations have been non-dimensionalized in terms of a lengthscale L, timescale T_D, and temperature scale T_m, the melt temperature. The coefficients $C(h)$, $K(h)$, κ_I, and τ are now non-dimensional. Also appearing are the density ratio $\delta = \rho_m/\rho_I$ and the Stefan number $C_I T_m / L_m$.

For a water-ice system we have $\delta \sim 0.9$ and $S \sim 0.46$. Typically, the heat diffusion timescale L^2/κ_I is much longer than the other timescales (infiltration timescale $\tau_i = L/K_s$, τ), so that to first approximation we can neglect heat diffusion in the ice. Then the behavior will be governed by the competition between infiltration and freezing. We can define a "Freeze number"

$$Fz = \frac{L}{K_s \tau} = \frac{\tau_i}{\tau} \tag{9}$$

such that small Fz means that infiltration dominates over freezing and large Fz is the opposite.

Figures 1 and 2 show sample 1-d calculations of infiltration into a region of size $L = 50$ km, with conductivity $K_s = 10^{-2}$ cm s^{-1}, so that the infiltration timescale $\tau_i = L/K_s = 5 \times 10^8$ s. Figure 1 shows an infiltration-dominated situation ($Fz = 5$), while Figure 2 shows one that is freeze-dominated ($Fz = 500$). For the smaller Fz case, melt infiltrates to the bottom of the domain before freezing and the ice temperature remains relatively low on that infiltration timescale of τ_i. For the high-Fz case, melt freezes as it infiltrates, so that melt fractions remain small except for a thin zone at the infiltration front. Latent heat from the melt readily warms the ice to near-melting temperatures.

Acknowledgments

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