CONVECTIVE INSTABILITY IN HORIZONTAL DECOMPACTION CHANNELTS INSIDE PLANETARY LITHOSPHERES. J. Schools[1] and L. G. J. Montési[1], [1]Department of Geology, University of Maryland, College Park (jschools@umd.edu; montesi@umd.edu).

Introduction: As melt travels upwards from the convecting mantle and through the lithosphere of planetary bodies, it likely becomes trapped below a horizontal layer (permeability barrier) at depths too deep to be readily tapped by dikes and fractures [1, 2]. Yet, melt must reach the surface to create the volcanoes and lava plains we observe throughout the solar system. We investigate convection in the melt rich, highly porous layer below the permeability barrier, known as the decompaction channel [3, 4, 5] as a potential mechanism for collecting melt and breaching the barrier.

Permeability barrier and decompaction channel evolution models have previously been developed in context of Earth's magmatism and plate motions, specifically dealing with mid-ocean ridges [e.g. 5, 6, 7, 8]. Under these conditions the barrier is sloped following the aging, cooling, and thickening of the lithosphere, and melts follow the sloping barrier towards the ridge axis. In other planetary bodies, such as Mars and Io [1, 2], that lack plate tectonics, permeability barriers are probably sub-horizontal. In this case, the melt lacks an obvious direction for melt ascension and remains in the decompaction channels. Here we discuss the possibility of convection in the decompaction channel and discuss its implications for allowing melt to traverse the barrier and possibly controlling the spacing of volcanic edifices.

Model: Using the finite element code ASPECT [9, 10] with melt migration extension [11], we model the formation and evolution of permeability barriers and decompaction channels in two dimensions, representative of a single plate planetary lithosphere. ASPECT with melt migration operates by solving the following equations for velocity, pressure, temperature, and porosity:

$$-\nabla\cdot[2\eta(\varepsilon(\mathbf{u}_s)-\tfrac{1}{3}(\nabla\cdot\mathbf{u}_s)1+\nabla p_f+\nabla p_c=\rho\mathbf{g} \quad (1)$$

$$\nabla\cdot\mathbf{u}_s-\nabla\cdot K_D\nabla p_f-K_D\nabla p_f\cdot\frac{\nabla\rho_f}{\rho_f}$$
$$=-\nabla\cdot\left(K_D\rho_f\mathbf{g}\right)+\Gamma\frac{\rho_s-\rho_f}{\rho_f\,\rho_s}-\frac{\phi}{\rho_f}\mathbf{u}_s\cdot\nabla\rho_s-K_D\mathbf{g}\cdot\nabla\rho_f \quad (2)$$

$$\nabla\cdot\mathbf{u}_s+\frac{p_c}{\xi}=0 \quad (3)$$

$$\frac{\partial\phi}{\partial t}+\mathbf{u}_s\cdot\nabla\phi=\frac{\Gamma}{\rho_s}+(1-\phi)(\nabla\cdot\mathbf{u}_s) \quad (4)$$

$$\bar{\rho}C_p\left(\frac{\partial T}{\partial t}+\mathbf{u}_s\cdot\nabla T\right)-\nabla\cdot k\nabla T=T\Delta S\Gamma \quad (5)$$

See [10, 11] for a more detailed explanation and derivation. The melting model follows the parameterization of [12] for dry peridotite. Crystallization is simplified as the reverse of melting.

Our model consists of a 2D box, 150 km in height and 160 km in width, representative of a younger, warmer Mars-like lithosphere. The top temperature is set to an average Martian surface temperature of 210 K. The bottom temperature is set to 1650 K, representing a young Mars. The initial temperature profile is set to:

$$T=T_s+\left\{(T_b-T_s)*\cos\left[\sin^{-1}\left(\frac{P-P_b}{P_b}\right)\right]^2\right\} \quad (6)$$

where T_s is the surface temperature of Mars, T_b is the temperature at the base of the lithosphere, P is the lithostatic pressure, and P_b is the pressure at the base of the lithosphere. Melt is simulated to enter the through the bottom of the system by setting the porosity of the bottom boundary to 99% and allowing liquid to flow in or out with pressure.

Results: In our nominal model, melt ascends, forming a permeability barrier near 101 km depth and a ~12 km thick decompaction channel underneath it. Starting at ~2.28 Myr, cold downwellings, consisting of both solid and trapped melt, descend from the top of the decompaction channel due to the crystallization of melt (Figure 1). Melting is at a maximum in the downwellings as the material is reheated. Upwellings bring hot melt and accompanying solid back to highly porosity pockets near the top of the decompaction channel. New drips form near the top of the channel as the newly arrived melt crystallizes.

Discussion:

Origin of convection: The potential for convection of a viscous fluid layer is expressed via the Rayleigh number, which a driving force to processes that dissipate it. For thermal convection, $\mathrm{Ra_{th}}=\frac{\alpha\rho_0 g\Delta T\,H^3}{\kappa\eta}$, where α is the coefficient of thermal expension, ρ_0 the reference density of the fluid, ΔT the temperature difference between the top and bottom of the layer, of thickness H, and κ and η are the thermal diffusivity and the viscosity. For our model parameters (Table 1), $\eta\sim10^{15}\mathrm{Pa}\cdot\mathrm{s}$ due to the high porosity in the decompaction channel, and $\mathrm{Ra_{th}}\sim383$, which is less than the critical Rayleigh number. As convection take places even if $\alpha=0$, we conclude that the forces driving convection are not purely thermal.

The density contrast between the fully-crystallized agregate at the top of the channel and the melt-rich aggregate provides a more important driver of convection. The associated Rayleigh number is $\mathrm{Ra_X}=\frac{(\rho_s-\rho_m)\phi g H^3}{\kappa\eta}$, where ϕ is the porosity in the decompaction channel. We observe $\phi\sim0.4$ and so $\mathrm{Ra_X}\sim1300$. Convection likely initiates due to the density differences between solid and melt.

Geological implications: Melt must be able to breach the permeability barrier and reach the surface. In this model, the permeability barrier maintains its depth overall but the highly permeable pockets at the top of the decompaction channel rise 200-300 meters as convection occurs. Melt is focused to these pockets where crystallization and latent heat release occur. It is possible that the heat released upon crystallization in the high melt content pockets provides a mechanism for melt to ascend past this the permeability barrier. In that case, melt will be focused over a length scale corresponding to the wavelength of the convective instability. A the onset of convection, that wavelength is $\sim 2.8H$. Therefore, magmatic upwellings take place every ~17 km. This scale will decrease as H decreases, which can happen if the heat flux, the grain size, or melt content decrease. Interestingly, the nearest neighbor spacing of small volcanoes on Syria Planum, Mars is 16.5 km [13], very close to our instability wavelength. These volcanoes were formed from the early Hesperian to early Amazonian, which is the time period our model approximates with its lithosphere thickness and thermal profile.

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Selected Parameters and Parameterizations	
$g = 3.7\ \mathrm{m/s^2}$	$k_\phi = k_{\phi_0}\phi^3(1-\phi)^2$
$\rho_{s,0} = 3000\ \mathrm{kg/m^3}$	$K_D = k_\phi/\eta_f$
$\rho_{f,0} = 2500\ \mathrm{kg/m^3}$	$\delta_c = \sqrt{\left(\frac{\xi + 4\eta}{3}\right)K_D}$
$\xi_0 = 10^{22}\ \mathrm{Pa\cdot s}$	$u_f = u_s - \frac{K_D}{\phi}(\nabla p_f - p_f g)$
$\eta_0 = 5\times 10^{20}\ \mathrm{Pa\cdot s}$	$\eta(\phi) = \eta_0 e^{\alpha_\phi(\phi-\phi_0)}$
$\alpha = 2\times 10^{-5}\ \mathrm{K^{-1}}$	$\rho = \rho_0[1-\alpha(T-T_a)]$

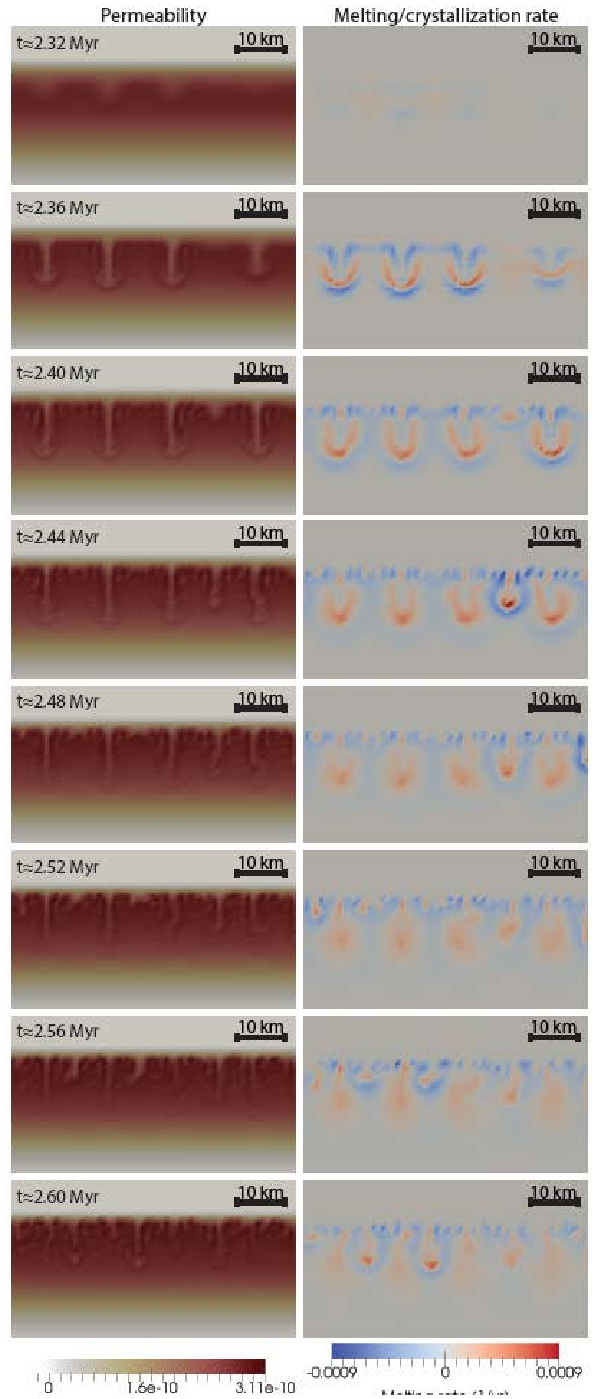

Figure 1: Evolution of convection in decompaction channel over approximately 280,000 years. Each row represents a time, separated by approximately 40,000 years. Column 1 shows the permeability of the model. Column 2 shows the melting and crys-tallization