GENERATION OF BARRIERS TO MELT TRANSPORT IN THE MARTIAN LITHOSPHERE.

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Introduction: Melt rises buoyantly from melt generation zones in the mantle toward the surface. However, melt ascent can be stopped if a permeability barrier forms as melt crystallizes in the cooler lithosphere. In that case, melt may be focused and form a central volcanic edifice. In this study we evaluate the conditions for which a permeability barrier may form in the lithosphere of Mars. Barriers due to the crystallization of olivine and clinopyroxene have likely formed throughout Mars' history, although melting and barrier formation would be restricted to mantle plumes over the last 1.5 billion years. Volcanism may require a heterogeneity in the lithosphere or relatively high strain rate in the mantle, which would link volcanic and tectonic areas.

Permeability Barriers: Within a warm mantle, a porous network with relatively high permeability allows melt to rise buoyantly [e.g., 1]. When this melt enters the colder lithosphere it begins to crystallize. If the crystallization is rapid enough the pore space will be clogged, drastically reducing matrix permeability and forming a permeability barrier [2,3]. At the Earth's mid-ocean ridges, a permeability barrier forms at the multiple saturation point of clinopyroxene and feldspar, which occurs when then melt temperature decreases to 1240°C+1.9*z*, where *z* is the depth in km [4].

Permeability barriers may halt the rise of melt to the surface of a planet. Mars displays a varied history of volcanism [5], therefore melt must reach the surface. By understanding the conditions which form permeability barriers we can address the nature of the lithosphere and mantle beneath volcanic centers.

Barrier Formation on Mars: We use the MELTS [6,7] and pMELTS [8] thermodynamic calculators with the alphaMELTS front-end interface [9] to determine melt compositions and the crystallization sequence of ascending melt throughout Mars's history. Starting with an anhydrous bulk silicate composition appropriate for Mars [10], we calculate the aggregate melt composition generated by fractional melting upon decompresssion along a mantle adiabat to a depth representing the base of the thermal lithosphere. The mantle potential temperature (MPT) and lithospheric thickness (*H*) are varied to create a continuum of possible melt percentages and compositions (Fig. 1a). The crystallization sequence of that melt is determined along a pressure-temperature path representing the geothermal gradient of the lithosphere from the termination of the melting calculation to the surface. The crystallizing mineral phases and crystallization rates are recorded. Crustal assimilation is not addressed in this model.

A permeability barrier will form where the compaction length δ_C is larger than the critical compaction length δ_C^* [3], or

$$\delta_C \geq \delta_C^* \text{ where } \delta_C^* = \left(\frac{dT}{dz}\frac{df}{dT}\right)^{-1} \text{ and } \delta_C = \sqrt{\frac{k_\theta\left(\zeta + \frac{4}{3}\eta\right)}{\mu}}$$

The compaction length depends on melt porosity ϕ and grain size *d* through the matrix permeability k_ϕ, whereas the matrix shear viscosity η and bulk viscosity $\zeta \sim \eta/\phi$ depend on strain rate $\dot{\varepsilon}$ and temperature. μ is the melt viscosity. In our reference model, ϕ=1%, d=3mm, $\dot{\varepsilon}=10^{-15}$ s^{-1} (Fig. 1c).

Permeability barriers typically develop as olivine, clinopyroxene, or both crystallize (Fig. 1c). Barriers due to the crystallization of feldspar are observed only for *H*<100 km and cool MPT, which is unlikely based on thermal evolution models [11]. Barriers due to the crystallization of nepheline are observed for large *H* and hot MPT, which is also unlikely. Additionally, nepheline saturates only late in the crystallization sequence, when little residual melt is present. Therefore, we consider that nepheline barriers do not strongly influence the style of volcanism.

No melt is generated in cases with large *H* and cool MPT. These conditions are expected to dominate over the last 1.5 billion years [11]. During that time, melt generation and formation of a permeability barrier are only possible in regions where the mantle is anomalously hot, as, for example, at mantle plumes.

The linear relation between temperature and depth at permeabilty barriers on Earth [4] is complicated on Mars by the change of MTP over time (Fig. 2). Previous work used a single Earth MPT and only identified barriers as crystallization peaks of feldspar and clinopyroxene. As we define permeability barriers instead with the above equations, a barrier may develop before or after the clinopyroxene peak as viscosity increases, leading to a deviation from the reference linear relation between depth and temperature.

Strain rate significantly influences the formation of permeability barriers. If the strain rate is reduced to 10^{-17} s^{-1} (Fig. 1b), a barrier forms upon crystallization of olivine for almost every *H* and MPT for which melting takes place in the mantle. If the strain rate is increased to 10^{-13} s^{-1} (Fig. 1d), no significant barrier forms except for *H* < 75 km, thinner than the modeled

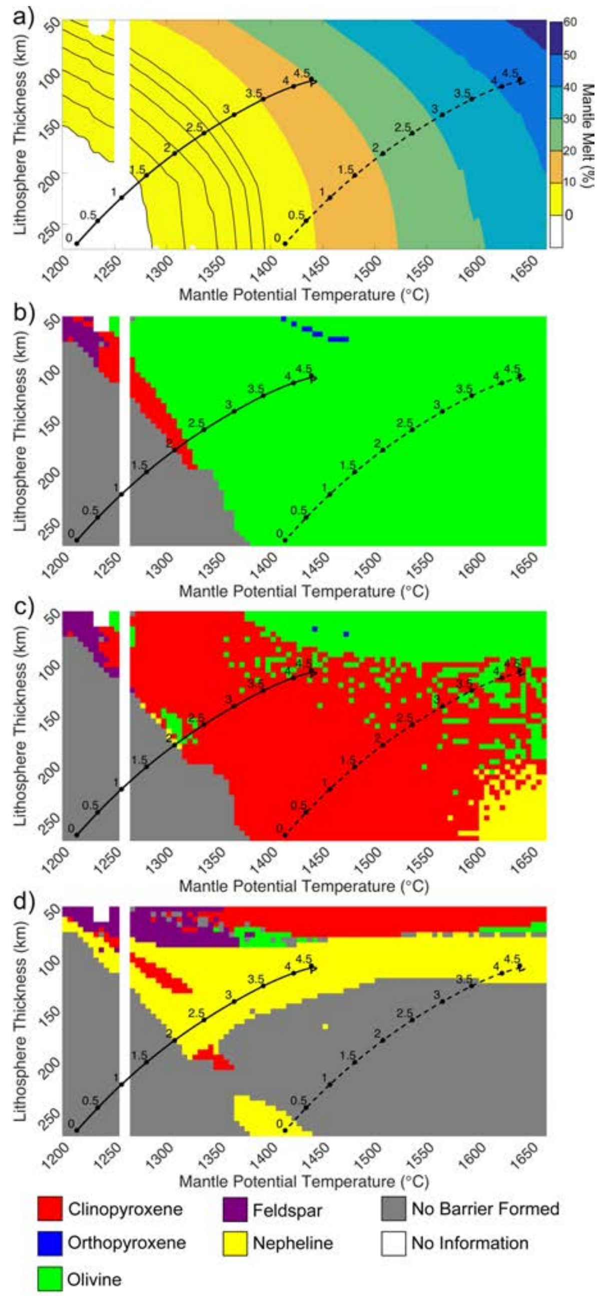

Figure 1: Results of barrier formation for an anhydrous bulk silicate Mars composition [10] with FMQ -2.5. a) Melting fraction of the mantle (black contour line intervals representing 0-5% melt). b-d) Main mineral crystallizing at the barrier assuming $\dot{\varepsilon}=10^{-17}$ s^{-1} (b), 10^{-15} s^{-1} (c), and 10^{-13} s^{-1} (d). The solid black line indicates the evolution of lithosphere thickness and MPT through time [11] and the dashed line the same evolution but adding 200°C to represent a mantle plume.

4.5 Ga average [11]. An increased strain rate scenario is plausible in mantle shear zones. Therefore, melt may be able to exploit actively deforming regions to reach the surface

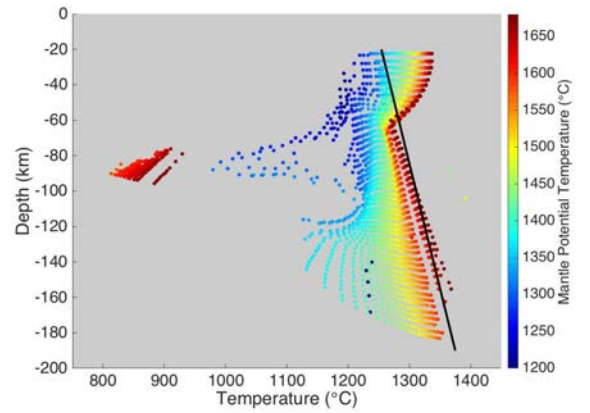

Figure 2: Temperature and depth at permeability barrier, color-coded for MPT. Conditions same as Fig. 1c. The black line represents the depth-temperature relationship of barrier formation at mid-ocean ridges [4], corrected for Martian gravity.

Implications: According to our thermodynamical calculations and the Martian thermal evolution model of Hauck et al. [11], there should be no melting, and therefore no active volcanism during the last 1.5 billion years of Martian history. Younger volcanic features [5] require an anomalously high temperature mantle, as may be expected at mantle plumes.

Our lower strain rate (10^{-17}-10^{-15} s^{-1}) models show barriers forming upon crystallization of olivine and clinopyroxene throughout Martian history. Based on studies of terrestrial mid-ocean ridges, melt that stalls at a permeability barrier can reach the surface only if it collects in specific regions related to heterogeneities in the lithosphere, or if it encounters a region of tectonic damage [12]. No significant barrier forms in models that assume higher strain rate. This possibly indicates that volcanism on Mars is associated with shear zones and other regions of active tectonic deformation.

References:

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