MODELING OF STEAM PRESSURE UNDER MARTIAN LAVA FLOWS. C. M. Dundas[1] and L. P. Keszthelyi[1], [1]Astrogeology Science Center, U. S. Geological Survey, 2255 N. Gemini Dr., Flagstaff, AZ 86001, USA (cdundas@usgs.gov).

Introduction: Mars has a long history of volcanic, aqueous and cryogenic processes. Interactions between lava and water or ice are expected, and many possible hydrovolcanic features have been reported [1-4]. The best-documented of these are rootless cones, formed when steam explodes through a lava flow and builds small cones on the flow surface [5]. Martian rootless cones were first proposed based on Viking imagery [e.g., 1,6]; higher resolution images have revealed better candidates [4, 7-10]. An important question is whether sufficient steam pressure can build up under a flow due to simple conductive heating of groundwater or ground ice; the alternative is that mixing of liquid lava and water or mud is required, generating a molten fuel-coolant interaction (MFCI) [11, 12]. The significance of this distinction is that ground ice on Mars is expected to lie beneath a lag of some thickness [e.g., 13], so a MFCI would require some combination of lava erosion of the substrate, flow of liquid water, or extremely recent emplacement of water or ice.

This is of special interest in Athabasca Valles, where rootless cones are found associated with an outflow channel coated by a thin layer of lava [9]. The lava flow was over 100 m thick at peak discharge and was fully turbulent, allowing the possibility that this 300 km long channel system was carved by lava, not water [14]. Alternatively, the channel system could be ancient and the young lava flow is a much later event [15] or the rising magma may have triggered the water flood that carved Athabasca Valles just before lava emplacement. These alternative formation scenarios predict different modes of lava-water interaction: erosion by lava allows for MFCI with deep ground ice; valley formation long before lava emplacement means that only atmospherically emplaced ground ice is available, though such ice is likely to be shallow if it occurs at all; and closely linked aqueous and lava floods allow for interaction with very shallow subsurface, or even surface, water. One of our major goals is to see if any of these models for Athabasca Valles can be eliminated due to the presence of rootless cones.

Model: We use two separate but coupled models to examine the accumulation of steam under lava flows. The first is a one-dimensional thermal model which is used to calculate the steam generation rate, and the second is a one-dimensional finite-difference model for gas transport through the porous lag following Darcy's Law.

The thermal model is a finite-difference model somewhat similar to that of [3]. The top boundary of the model is the lava flow core, followed by a desiccated lag of varying thickness and then a layer of ice-cemented regolith. There are multiple model layers in each material layer. To simplify the thermal model, we maintain the upper boundary of the model at 1070 °C (the crust boundary temperature of [16]) and add layers to the top over time to simulate the development of a lower crust rather than directly model the cooling and crystallization of the lava flow. The growth rate of this crust is parameterized to be one-tenth that of the upper crust growth rate determined by [16] for pahoehoe flows in Hawai'i. This can be improved in the future with a more complete numerical model for lava flow cooling similar to that of [17], but this model is adequate to quantify the basic behavior of the system.

Gas transport is modeled using Darcy's Law with gas added uniformly to each cell at each timestep. The outer edge of the model is held at atmospheric pressure, estimated at 800 Pa, and there is no flow across the flow center (assumed symmetric). The system is assumed to have uniform porous lag thickness, temperature and permeability, and the pressure in each cell is assumed to be constant, with no vertical gradient.

In order for steam pressure to generate an explosion, the pressure must be high enough to overcome both the weight and the yield strength of the lava [8]. Lava flows 1-50 m thick with a bulk density of 2200 kg m^{-3} exert an overburden pressure of 8×10^3 – 4×10^5 Pa, while the effective strength of the lava is likely ≤ 10^6 Pa [8]. We track the pressure buildup until it reaches 1.6×10^6 Pa.

The standard model inputs considered are generally conservative. We assume a permeability of 10^{-11} m^2, which is near the high end of a range of Martian soil analogs measured by [18], and a lava flow half-width of 500 m. For the shallower ice tables considered, these parameters turn out to not be critical since explosive pressures are reached before the low pressures at the model edge can greatly affect the center.

Results: In general, we find that explosive pressures are readily reached for ice table depths of a few decimeters to over a meter. The time to reach those pressures varies greatly with the ice table depth, both because more steam is needed to pressurize a thicker lag and because steam generation is slower and begins later for deeper ice. Pressure generally builds rapidly once steam generation begins.

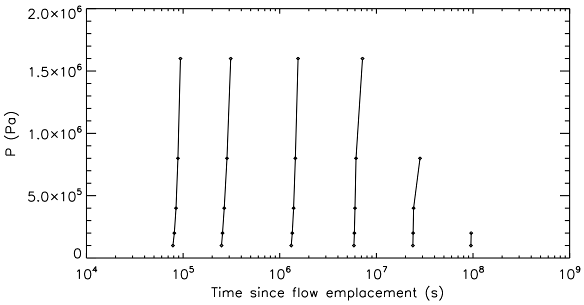

Figure 1: Peak pressure during steam buildup as a function of time since lava emplacement for ice table depths of (left to right) 0.2, 0.4, 0.8, 1.6, 3.2, and 6.4 m, assuming a permeability of 10^{-11} m^2 and a lava flow half-width of 500 m.

Discussion: These results demonstrate that it is possible to build up explosive pressures through passive heat conduction. This does not exclude the possibility of MFCIs. However, they may not be required to initiate rootless eruptions. We note that some of the lag material may be entrained in the initial eruption, which could explain the observation of mixing of lava and sediments near the beginning of eruptions [12].

After an initial explosion, it is likely that there will be a steep pressure gradient around the eruption site that will drive inward flow of both steam and water. This will facilitate the accumulation of steam for further eruptions and could also lead to MFCIs as both fresh lava and water flow to the eruption site.

For thin dry lags and thick lava flows, the energy stored by pressurized steam within the lag pore space may be inadequate to raise all of the overlying lava column above the lava surface. Such cases might only lead to the formation of spiracles.

These results can also be compared with observations of Martian rootless cones. They are observed to occasionally form in rings, presumed to lie over the rims of buried craters, and smaller cones are observed on high topography where the lava flow was thinner [4, 9, 10]. Both of these are consistent with steam pressurization, since eruptions will preferentially occur where the lava overburden is thin, but may be less energetic precisely because lower pressures are required. Variations in lag thickness or permeability could also contribute to localization of eruptions.

Rootless craters on the Athabasca Valles lava flow are inferred to have formed ~3-24 days after the development of a stable crust on the flow [14]; this was likely some time after the initial emplacement of the flow, but the overall eruption duration is thought to be a few months. This is comparable to the time at which explosive pressures are reached for lava depths of 0.4-0.8 m for the conditions in Fig. 1.

We have separated the thermal and gas-transport models for simplicity; however, this does lead us to omit some potentially relevant effects. We ignore the possibility of water movement beneath the lava flow, driven by gradients in steam pressure. Near the flow center the pressure gradient is low, so this can be neglected. However, near the flow boundaries the gradient may be steep enough to force meltwater out from under the flow. We have also omitted the pressure dependence of the water boiling point, because we do not track the gas pressure (which varies in space and time) in the thermal model. Instead, we have used the value for one bar at all times. The rise of the boiling point at higher pressures would reduce the steam generation rate but is unlikely to fundamentally change the results, only slow the time to reach peak pressures, except for cases that barely reach 1.6 MPa. However, the slopes of the pressure rise curves in Fig. 1 may be substantially less steep. Model runs with an elevated boiling point support this suggestion. Smoothing of the steam flux to remove effects from finite-differencing (similar to [17]) may also offset curves somewhat.

Conclusions: Steam is able to build to explosive pressures under lava flows through passive heat conduction with an ice table at a depth up to a few meters, depending on the strength and thickness of the flow. This enables rootless cones to initiate without MFCIs, although such interactions are possible and may be more likely after an initial explosion. Eruptions due to steam buildup should preferentially occur near flow centers and over buried topographic highs. At this point we have not been able to eliminate any of the formation models for Athabasca Valles, but a more detailed comparison of the observations and model results is still needed.

References: [1] Allen C. C. (1979) Ph. D. dissertation, U. Arizona. [2] Mouginis-Mark P. J. (1985) *Icarus*, 64, 265-284. [3] Squyres S. W. et al. (1987) *Icarus*, 70, 385-408. [4] Keszthelyi L. P. et al. (2010) *Icarus*, 205, 211-229. [5] Thorarinsson S. (1953) *Bull. Volc.*, 14, 3-44. [6] Frey H. et al. (1979) *JGR*, 84, 8075-8086. [7] Lanagan P. D. et al. (2001) *GRL*, 28, 2365-2367. [8] Greeley R. and Fagents S. A. (2001) *JGR*, 106, 20527-20546. [9] Jaeger W. L. et al. (2007) *Science*, 317, 1709-1711. [10] Hamilton C. W. et al. (2011) *JGR*, 116, E03004. [11] Fagents S. A. and Thordarson T. (2007) in *The Geology of Mars*, Cambridge Univ. Press. [12] Hamilton C. W. et al. (2010) *Bull. Volc.*, 72, 449-467. [13] Leighton R. B. and Murray B. C. (1966) *Science*, 153, 136-144. [14] Jaeger W. L. et al. (2010) *Icarus*, 205, 230-243. [15] McEwen et al. (2012), this conference. [16] Hon K. et al. (1994) *GSA Bull.*, 106, 351-370. [17] Milazzo M. P. et al. (2001) *JGR*, 106, 33121-33127. [18] Sizemore H. G. and Mellon M. T. (2008) *Icarus*, 197, 606-620.