GEOPHYSICAL CONTROLS ON MARTIAN POST-IMPACT HYDROTHERMAL SYSTEMS.

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Introduction: Current surface temperatures and pressures maintain a Martian cryosphere that is at least hundreds of meters thick [1]. However, it has long been recognized that the energy delivered by bolide impacts are likely to melt and/or vaporize the subsurface ice [2,3,4]. Expanding on ideas present by [5], [6] postulated that large impactors of ~100 km or more in diameter could liberate enough subsurface water to induce transient greenhouses.

Smaller impacts deliver energy locally through shock heating and elevated geothermal fluxes that cause long-lived subsurface fluid circulation [4,7,8,9,10]. These post-impact hydrothermal (PIH) systems likely bring water close enough to the surface to leave an observable signature.

Here we present analytic estimates on the effects of two principal controls on PIH systems: permeability and heat sources. These and other analytic approximations will be used to motivate and benchmark a more sophisticated analysis of PIH systems and their effect on the Martian cryosphere using numerical simulations.

Geophysical Controls: The evolution of subsurface ice and water in response to PIH systems is a complex process dependent upon heat sources and permeability. The impact event itself is the most important initial condition; the energy delivered by the impact and the redistribution of shock-heated material sets the stage for any subsequent hydrothermal activity.

Heat Sources: There are three heat sources that contribute to the initial temperature profile and drive the PIH system: impact induced shock heating, the melt sheet pooled on the crater floor, and the uplifted geotherm (see figures 1 and 2). The distribution of shock heating produced by an impactor can be estimated by combining an analytic expression for the change in internal energy as a function of distance from the impact point with pi scaling arguments that quantify the morphological change from the transient to final crater (cf. [11]). The geothermal heat flux is a key parameter governing deep flow in PIH systems and the distribution of hypothetical aquifers beneath the cryosphere [1]. Although no direct measurements exist, estimates for the geothermal heat flux cluster around ~30 mW m^{-2} [12,13,14].

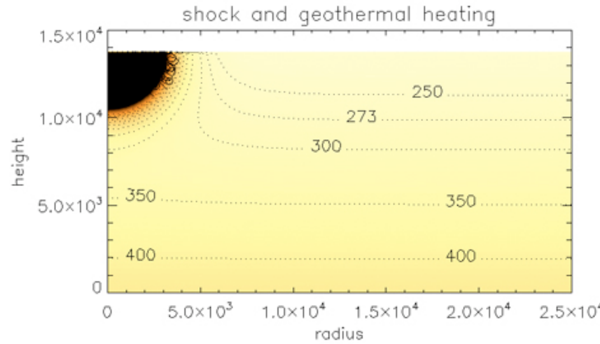

Fig 1. Temperature contours of shock heating generated by an impactor 3.6 km in diameter with a velocity of 10 km s^{-1} and a density of 2600 kg m^{-3} superimposed on a background geothermal temperature profile established by a geothermal flux of 30 mW m^{-2}, a thermal conductivity of 2 W m^{-1} K^{-1} and a surface temperature of 215 K.

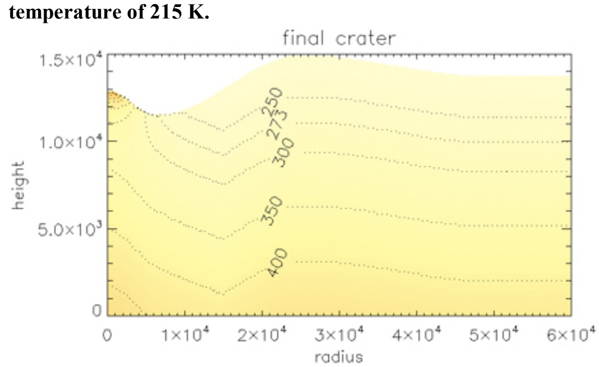

Fig. 2. A rough approximation of a post-impact temperature profile for a 50 km diameter final crater. The impactor described in Fig 1 excavated material to a depth 3.7 km at the crater's center and created a transient crater 11.1 km deep that rebounded isostatically. If water is present, it will be in the ice phase above the 273 K contour line. The residual shock heating concentrated under the central peak and the up-warped crust are two heat sources that will drive hydrothermal circulation. A melt sheet was not included in this approximation.

An order of magnitude estimate of the thermal conductive time can be calculated analytically:

$$t \sim \frac{l^2}{\kappa} \tag{1}$$

where l is a characteristic length scale, and κ is the thermal diffusivity. An l of 1 km a $\kappa \sim 10^{-6}$ (i.e. basalt) yields a conductive cooling time of ~300,000 yrs.

Permeability: The permeability characterizes the resistance of the porous medium to flow through it and has a profound effect on system lifetime. The permeability of the upper Martian crust is unknown. Lower permeabilities of 10^{-3} to 10^{-1} that decay exponentially were adopted by [10], while [15] chose a higher permeability of 10^3 darcys typical of young terrestrial basalts with horizontal fractures.

Assuming 1D steady state flow driven by the buoyancy of hot water we derive the darcy velocity (v) of a rising plume and hence a heat flux (q) (Figure 3) and system cooling time approximation (t) (Figure 4) as follows (cf. [16]):

$$v = \frac{k\alpha\rho g}{\mu}\left(T_r - T_0\right) \tag{2}$$

$$q = v\rho C_p\left(T_r - T_0\right) \tag{3}$$

$$t = \frac{\mu l}{k\alpha\rho g}\left(\frac{1}{\Delta T_0}\right) \tag{4}$$

where α, ρ, and μ are the thermal expansivity, density, and viscosity of water, k is the average permeability, l is a characteristic length scale, chosen to be 1 km, g is gravity, T_r and T_0 are the reservoir and surface temperatures, and ΔT is the difference between the system reservoir and surface temperature.

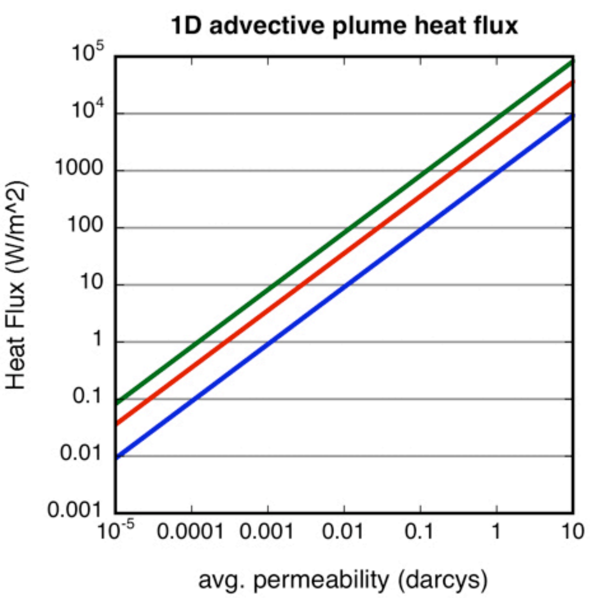

Fig. 3: Heat flux as a function of permeability (1 darcy = 0.987×10^{-12} m^2) for three ΔTs: 90 (blue), 180 (red) and 270 (green)

Future Work: PIH systems will be modeled as a multi-phase, convective flow problem through a saturated, heterogeneous, porous medium in which both advection and conduction will contribute to temporal and spatial changes in temperature. The code we will employ is MAGHNUM (Mars Global Hydrology Numerical Model), developed by Bryan Travis. MAGHNUM solves governing equations for conservation of mass and energy transport. Unlike codes used in previous studies, MAGHNUM includes ice freezing and is therefore an especially capable tool to investigate PIH systems on Mars.

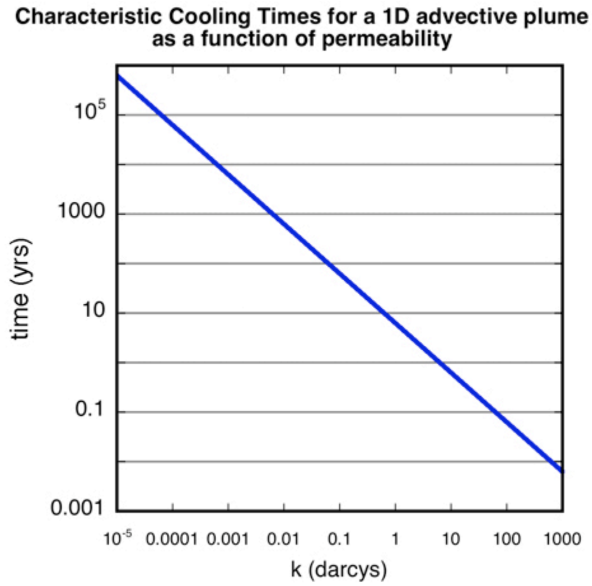

Fig 4. The time required for an advective plume to equilibrate with the background geotherm from an initial ΔT of 180K decreases with permeability (1 darcy = 0.987×10^{-12} m^2). Low permeabilities yield longer lived but less vigorous PIH systems.

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