DYNAMIC MODELING OF MARTIAN PALEOLAKE STABILITY. E. G. Rivera-Valentin[1], R. Ulrich[2], V. F. Chevrier[1], T. S. Altheide[1], J. J. Wray[3], [1]Arkansas Center for Space and Planetary Sciences (eriverav@uark.edu), [2]Dept. of Chemical Engineering, University of Arkansas, [3]Dept. of Astronomy, Cornell University.

Introduction: Recent studies have found positive identification of the location of paleolakes on the surface of Mars via observations of standlines [1] and crater floor polygons too large to be formed by only thermal processes [2]. By analyzing CRISM spectral data, Wray *et al.* have also shown the existence of polyhydrated sulfates and salts on the floor and inner walls of Columbus crater, Mars (29.8°S, 166.1°W) [3].

With such evidence in mind, we conducted an in-depth stability study of paleolakes on Mars by creating a finite element model that couples heat and mass transfer. Our model is written in MatLab and incorporates multiple simultaneously running processes such as sublimation, freezing, diffusion-advection, and heat transfer. Our goal is to more precisely analyze the lifespan of a paleolake while observing such elements as changes in the activity of water while freezing, mineral precipitation, post-freezing residue brine deposits and the effects of a soil layer on the ice cap surface.

Methods: We modeled a 1 m^2 column in the lake that is not in direct contact with any of the crater walls, thus we can assume an adiabatic boundary between the column and identical columns around it. While it has been suggested that if the initial temperature of the emplaced fluid is sufficiently elevated, ice-cap synthesis can be avoided for ~3-8yrs [4], it is assumed that upon fluid deposition the lake immediately forms an ice cap and thus is initially at its freezing point. We also assume the lake is vertically well mixed.

Heat Flux Model: We primarily use the equation set proposed by multiple authors to model the heat flux incident on the martian surface to reproduce diurnal temperature change [5,6,7]. We take into account the diffusion of the direct solar beam, the indirect solar illumination due to scattering, and thermal emission of the atmosphere as follows:

$$Q_{DB} = I_{sun}(1-A)\cos(z)T(z,\tau) \quad (1)$$

$$Q_{scat} = \left(I_{sun}\left(1-T(z,\tau)\right) - I_{ab}\right)(1-A)f_{scat} \quad (2)$$

$$Q_{atm} = I_{sun}(1-A)\varepsilon f_{atm}\cos(\delta-\varphi) \quad (3)$$

where I_{sun} is the average solar flux received at Mars' orbit, I_{ab} is the amount of solar flux absorbed by the atmosphere, A is albedo, z the zenith angle, τ the opacity of the atmosphere, $T(z,\tau)$ the transmission coefficient [8], f_{scat} and f_{atm} are fractions of flux provided by Schmidt *et al*. (2009) [5], ε is the atmospheric thermal emissivity, δ is the solar declination, and φ the latitude.

We also include a moderate geothermal heat flux assigned as *30 mW/m²* [9]. Heat flux loss mechanisms are derived from surface blackbody radiation and evaporative cooling. Soil thermal properties are derived from Phoenix data [10].

Sublimation Model: To account for both the cases when there exists clean ice and a soil surface layer above the ice, we apply a rate limiting sublimation equation that takes into account both Ingersoll's flux equation [11] and diffusion advection [12] written as:

$$J_{tot} = \left(\frac{1}{J_{Ing}} + \frac{1}{J_{DA}}\right)^{-1} \quad (4)$$

where

$$J_{Ing} = (0.17)D_{H2O/CO2}\,a_{H2O}\,\Delta\eta\left[\frac{\left(\frac{\Delta\rho}{\rho}\right)g}{\upsilon^2}\right]^{1/3} \quad (5)$$

and

$$J_{DA} = D_{H2O/CO2}\frac{P}{RT}\frac{1}{L}\ln\left(\frac{1-y_{atm}}{1-y_{sat}}\right) \quad (6)$$

Here $D_{H2O/CO2}$ is the diffusion coefficient of H_2O through CO_2, a_{H2O} is water activity, $\Delta\eta$ is the water vapor density gradient between the surface and the atmosphere, $\Delta\rho/\rho$ is the surface to atmosphere relative density difference, g is gravity, ν the kinematic viscosity of CO_2, P/RT is the total ideal gas concentration, L the soil depth, and y_{atm} and y_{sat} are the mole fraction of water vapor in the atmosphere and at saturation.

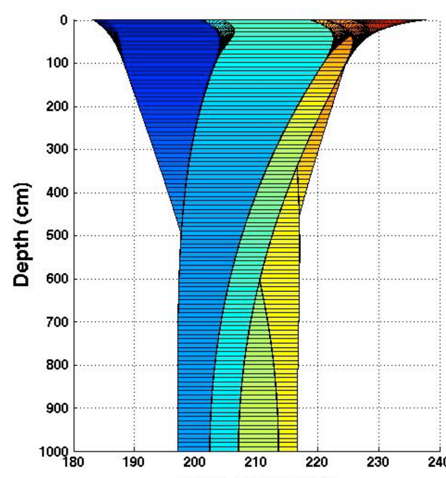

Fig. 1: Temperature with depth profile across a year for an ice lake on Mars. Colors represent temperature with red being the warmest and blue being the coolest. Each vertical curve is one day recorded at intervals of 60° of Ls.

Freezing Model: We consider the flux of liquid H_2O into the solid state as follows:

$$J_{Frz} = \frac{k_{ice}}{\Delta H_{fus}} \frac{\Delta T}{\Delta z} \quad (7)$$

where k_{ice} is the thermal conductivity of ice, ΔH_{fus} is the enthalpy of fusion, and ΔT and Δz are changes in temperature and depth at the solid-liquid interface. The heat flux due to freezing is accounted for at the solid-liquid interface. It is during freezing that the removal of H_2O from the solution increases the salt concentration thus lowering the activity of water and depressing the freezing point of the solution. We specifically observe when the freezing point of the solution is much lower than the current temperature at the solid-liquid interface for post-freezing residue brine deposits.

Results: We tested our thermal model against a wide range of conditions and have shown that it accurately predicts martian surface temperatures [13]. For our preliminary results, we assume dry martian atmospheric conditions, employ the dry adiabatic lapse rate to calculate the ambient temperature, observe the temperature fluctuations at latitude 29.8°S, and assume a completely frozen body.

In Figure 1, we plot outputted temperatures at multiple ice depths recorded at intervals of 60° of Ls for one year. The area between data is filled to show expected temperature variations across the year and colored to represent temperature magnitude.

Figure 2 plots the change in height of the frozen lake across the year between summer seasons. Multiple lines are shown representing differing amounts of soil deposits. The inflections seen at around 360° and 180° are the flux change during the winter season. It is seen that with increasing surface soil deposit, the change in height tends towards a linear relationship due to the decrease of temperature variation at the ice surface.

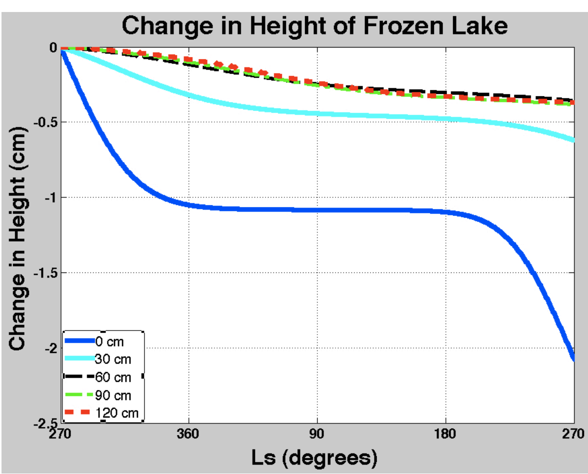

Fig. 2: Predicted decrease in frozen lake height across the year with varying amounts of surface soil deposit.

Figure 3 shows the effect of a soil deposit on the lake's life-span. This is found after studying the life expectancy of a frozen lake with 0, 30, 60, 90, and 120 cm of surface soil deposit. Life expectancy is seen to slightly drop around the annual skin depth due to smaller temperature variations of about 10 K centered around 220 K.

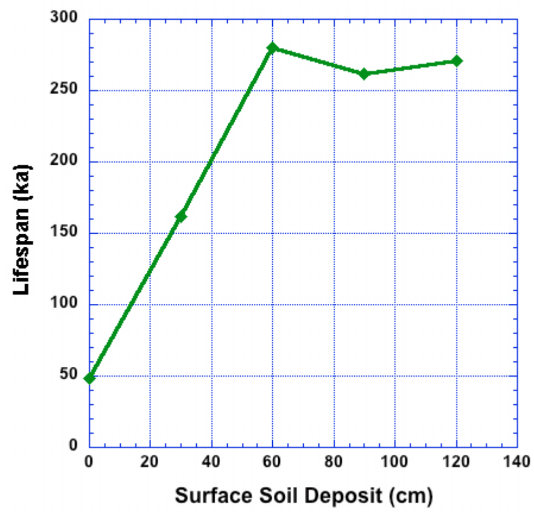

Fig. 3: Effect of surface sediment deposit on the expected lifespan of a frozen lake at latitude 29.8°S.

Conclusion: Our projected lifespans for a 1 km martian frozen paleolake correlates well with previous studies [4, 14]. Preliminary analysis show that the lifespan of the paleolake will increase with depth of overlying regolith until the annual skin-depth is approached; at which point, the life expectancy drops slightly and levels off. The results of our fully integrated model will be presented at the conference.

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