FIRN DENSIFICATION ON LATE NOACHIAN MARS: IMPLICATIONS FOR ICE SHEET FORMATION AND THERMAL CHARACTERISTICS. J. P. Cassanelli[1] and J. W. Head[1], [1]Brown University Department of Geological Sciences, Providence, RI 02912 USA (James_Cassanelli@Brown.edu).

Introduction: Recent Global Climate Model (GCM) simulations of an early CO_2 martian atmosphere and Late Noachian climate predict that beyond a fraction of a bar, atmosphere-surface thermal coupling occurs and an adiabatic cooling effect results in the deposition and accumulation of snow in high areas, particularly across the southern uplands of Mars [1,2]. Under these conditions water ice deposits will remain stable in the highlands, leading to a Late Noachian climate characterized by an "icy highlands" scenario [2].

In this study, we assess the nature of the snow accumulation and densification processes and the implications for potential mechanisms of melting of the icy highlands regions. We assume that the supply of water is governed by the surface water reservoir [3] and thus that the deposition of snow across the highlands is a supply-limited process [4,5]; we adopt a surface reservoir 2 x the amount of water-ice currently available in the polar/near-surface reservoirs (~60 m GEL). Given the supply limitation of water-ice, calculations suggest that regional ice accumulation does not exceed ~200-meters in thickness [3,4]. We adopt net accumulation rates predicted for more recent episodes of mid-latitude glaciation [6].

Previous work has analyzed firn densification of the north polar deposits on Mars to determine if the polar deposits might contain a past climate record in the form of entrained gas bubbles [7]. Here we seek to analyze the general process by which snow densifies into firn and ice and the effect this will have on melting and runoff under Late Noachian conditions [5]. We define snow as having densities below 350 kg/m^3; firn, densities between 350 and 830 kg/m^3; and ice, densities greater than 830 kg/m^3.

The densification process has implications for the overall formation and thermal properties of the resulting deposits. Characterization of the thermal properties of the ice deposits places important constraints on the susceptibility of the deposits to punctuated heating from both top-down (ideal insolation geometry, impacts, volcanism) and bottom-up (increased geothermal heat flux, excess ice accumulation) sources [5].

Methods: To simulate the densification process we have implemented the following relation [8]:

$$\frac{d\rho}{dz} = \frac{f_s \rho^2}{\rho_i}\frac{1}{b}\exp\left(\frac{-Q}{RT}\right)\left[\left[\frac{\rho_i}{\rho} - 1\right]P_*\right]^n \qquad (1)$$

This equation relates the change in density with depth (dρ/dz) to the physical parameters which control the densification process. This model was selected for the analysis of martian ice deposits because it is physically derived from the governing processes of densification, and does not rely on empirical relations or calibrations. The model as it appears in equation (1), however, does not account for the enhancement of grain settling caused by sublimation. Given the generally large sublimation rates predicted to occur on Mars [6] we have incorporated a sublimation term to account for this effect: $\dot{S}e^{(\frac{550}{\rho}-1)}$. The value of the sublimation coefficient ($\dot{S}$) was determined by calibration with empirical data from Siple Dome, Antarctica [9]. The model was parameterized according to the Siple Dome conditions, and the sublimation coefficient was adjusted until a best fit was achieved (R^2~0.99). Once calibrated, the model was parameterized for Late Noachian conditions using results from GCM simulations [1,2], and used to generate density versus depth curves.

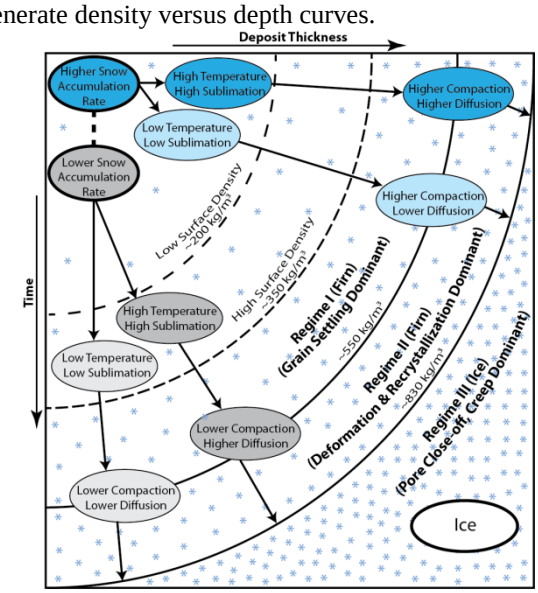

Figure 1. Conceptual diagram outlining the densification process in response to differing climate conditions. Within Regime 1 densification proceeds primarily by grain sliding until the availability of space in-between individual grains is reduced such that further densification cannot occur without grain deformation. Densification in Regime 2 continues through grain deformation and vapor diffusion redistributing ice at grain contacts (recrystallization). Regime 3 is entered when the pores in-between ice grains become isolated, further densification occurs by creep compressing the individual pore spaces.

Results: Figure 2 displays the analysis of the densification process under the outlined conditions. Regional snow accumulation occurs at a rate of approximately 10 mm/yr and continues until the ice reservoir becomes depleted. The analysis indicates that after deposition, less than half of the deposit is ice due to the low temperatures and low compaction forces arising from the weaker gravitational acceleration on Mars.

After deposition, however, the snow and firn will continue to densify through grain settling, deformation, and creep (Regime 2). If deposition ceases due to supply limitation, after 1 Ma nearly 90% of the deposit will have transitioned into ice. The densification will also result in a decrease in the overall thickness of the deposit by approximately 10% after 100 Ka, 16% after 1 Ma and 20% after 10 Ma. If conditions allowing regional ice deposits to remain stable in the highlands persist for between 10-100 Ma, most of the deposits will have transitioned to ice,

leaving an ice deposit with overlying firn with firn-to-ice transition depths of ~10-20 m.

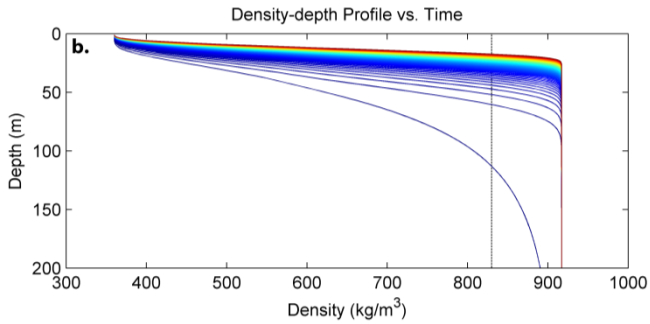

Figure 2. Density versus depth curves of a 200 m snow accumulation under Late Noachian conditions from time of deposition to 1 Myr afterwards (warmer colors correspond to increasing time).

Response to Heating: The resulting density profile of the ice deposits has implications for the response of the deposits to both *top-down* and *bottom-up* transient heating events. The amount of energy required to melt a unit volume of firn or ice is directly proportional to its density ($Q \sim 334\ J/cm^3 * \rho/\rho_i$). Therefore, a volume of firn with a density equal to half that of pure ice will take exactly half as much energy to melt as an equivalent volume of ice.

With respect to *top-down melting* this has two important implications: 1) The deposits will be more susceptible to top-down melting because it will require less energy to initiate and sustain melting of the less dense upper layers and because the deposits considered here will not be thick enough to cause a large rise in temperatures at the base. 2) As the deposits become older the firn-to-ice transition trends towards the surface as shown in Figure 2. This leads to a reduction in the amount of overlying snow and firn and increases the amount of energy required to melt down to a given depth within the deposit. Top-down heating will first cause melting of the surficial layers (snow and firn) of the ice sheet. In order to melt the uppermost meter of the modeled ice sheet a punctuated heating event would need to raise the surface temperature to 273K for a minimum total of approximately 18 days. Melting of one square kilometer of an ice sheet with a nominal surface density of 360 kg/m^3 to a depth of one meter will result in the generation of 360,000 m^3 of melt water.

Meltwater generated at the surface will begin to migrate down into the ice sheet by two major pathways: 1) fractures, crevasses, and moulins 2) percolation through the overlying permeable snow and firn. Meltwater that migrates downward through fractures, crevasses and moulins will be able to penetrate further into the ice sheet. This meltwater will then refreeze once a certain depth is reached because the ice sheet remains cold-based through the punctuated *top-down* heating event. When the meltwater refreezes it will release latent heat and warm the surrounding ice, possibly enhancing local densification. Alternatively, it is possible for some water to drain through fractures, crevasses and moulins to coalesce in englacial tunnels which might drain to periglacial areas.

Meltwater that percolates through the snow and firn will propagate downwards until it refreezes or encounters an impermeable layer of firn or ice. If the infiltrating meltwater refreezes, it will deposit latent heat into the surrounding firn and create ice layers, effectively enhancing the densification. If the meltwater does not refreeze it can enter storage in the porous firn layers, akin to typical terrestrial aquifers. Acting as an aquifer, the snow and firn layer can store significant quantities of meltwater and may also delay refreezing [10]. Given the low temperatures and gravitational acceleration on Mars it is likely that a snow and firn layer would be present across most or all of the modeled ice sheet. As meltwater is introduced the firn layer will gradually become saturated and the water table will rise towards the surface of the ice sheet. In areas of gentle surface slope the water table can intersect the surface and lead to the formation of meltwater ponds, particularly in areas of thin firn cover. Once formed, the melt ponds enhance local melting through more efficient absorption of incident solar energy. The melt ponds can then drain down into or off the ice sheet by steady discharge or episodic flooding [3].

The density profile of the deposit also affects *bottom-up heating* in two important ways: 1) As described for the surficial layers, older deposits will be denser and will require more energy input to the bottom of the deposit to cause melting. 2) Ice becomes more thermally conductive as the density increases. This means that a layer of ice provides less insulation than an equivalent layer of firn (the insulation of the ice deposit is dependent on the integral of the thermal conductivity across the ice sheet). Thus, the older and denser the deposit becomes, the less insulation it will provide and the more heat it will require for basal melting. Alternatively, if *top-down heating* were to take place, a denser ice sheet would transmit the increased surface temperatures to the base of the ice sheet much more quickly. Under nominal Late Noachian conditions, basal melting of a 200 m ice deposit would require an extremely high geothermal heat flux, approximately 700 mW/m^2, to achieve basal melting. These results suggest that melting will most readily be achieved by top-down heating events, and by bottom-up heating in conjunction with some degree of top-down heating.

Accumulation Rate (m/yr)	Geothermal Heat Flux (mW/m^2)		
	45	75	100
0.0	3.2 (269.8)	5.4 (267.6)	7.1 (265.9)
0.01	3.1 (269.9)	5.3 (267.7)	7.0 (266.0)

Table 1. Temperature increases (°C) across a 200 m ice deposit for a range of predicted Noachian geothermal heat fluxes (45-75mW/m^2) [11] and a volcanic geothermal heat flux (100 mW/m^2) [12] at zero accumulation and at an average Noachian accumulation rate [2] (required mean annual surface temperature for basal melting).

References: [1] Forget et al. (2013) *Icarus, 222,* 81-99. [2] Wordsworth et al. (2013) *Icarus, 222,* 1-19. [3] M. Carr, J. Head (2014) *LPSC 45,* Abstract #1427. [4] J. Fastook, J. Head (2014) *LPSC 45,* Abstract #1115. [5] J. Head et al. (2014) *LPSC 45,* Abstract #1412. [6] Madeleine et al. (2009) *Icarus, 203,* 390-405. [7] Arthern, R.J. et al. (2000) *Icarus, 144,* 367-381. [8] Cuffey, K.M., and Patterson, W.S.B., *The Physics of Glaciers* (4th ed.), Elsevier, p. 22-23. [9] Hawley et al., (2004) *J. Glaciology, 50,* 170. [10] Forster et al. (2013) *Nature Geoscience,* (DOI 10.1038/ngeo2043) [11] J. Fastook *et al*., *Icarus,* 219, 25 (2012). [12] Fasset C.I. and Head J.W. (2007) *Icarus, 189,* 118-135.