MARS SUBSURFACE WARMING AT LOW OBLIQUITY.
Stephen E. Wood[1] and Stephen D. Griffiths[2],
[1]Dept of Atmospheric Sciences, Univ. of Washington, 408 ATG Bldg., Seattle, WA 98195-1640, USA, sewood@atmos.washington.edu, [2] Department of Physics, University of Toronto, 60 St. George Street, Toronto, ON, M5S 1A7, Canada, sdg@atmosp.physics.utoronto.ca

Summary: We present a modeling study of a mechanism that has not previously been considered but is likely to have generated significant subsurface warming during the periodic intervals when Mars' obliquity was lower than 25°. Orbital dynamics calculations show that Mars' obliquity, which is currently 25°, oscillates between 10° and 45° - with a dominant periodicity of ~120,000 years and a modulation period of ~1.3 million years - due to long-term perturbations by the other planets [1]. The present Martian atmosphere is 95% CO_2 with a mean surface pressure of 700 Pa, but model calculations [2] show that it could drop to as low as 30 Pa at low obliquity because the global surface pressure would be controlled by the annual-average temperature of the perennial CO_2 ice at the poles [3]. At such low pressures, the thermal conductivity of a porous regolith can be significantly reduced as the mean free path of gas molecules approaches the size of pore spaces [4]. This decreased conductivity leads to increased subsurface temperatures as the geothermal gradient steepens to maintain a constant internal heat flux (estimated to be 0.030 W m^{-2}). We have performed model simulations of the resulting time evolution of subsurface temperatures and find that increases of 20-30 K are possible at latitudes and depths where ground ice may still be present. This could explain many of the geomorphological features attributed to liquid water without invoking a thicker atmosphere or increased geothermal heat flow.

Thermal Conductivity of Martian Regolith: In order to estimate the thermal conductivity of Martian regolith over a wide range of pressures and particle sizes, we have developed a physical model based on a recent analysis of the effective thermal conductivity bounds of isotropic, porous materials [5]. According to this analysis, a planetary regolith would be classified as an "external porosity" material and have an effective thermal conductivity (k_{eff}) bounded above by Equation (1) and below by Equation (2):

(1) $k_{max} = \frac{1}{4}\{(3\phi-1)k_a + [3(1-\phi)-1]k_s + \mathrm{sqrt}([(3\phi-1)k_a + (3\{1-\phi\}-1)k_s]^2 + 8k_ak_s)\}$

(2) $k_{min} = k_a[2k_a+k_s-2(k_a-k_s)(1-\phi)]/[2k_a+k_s+(k_a-k_s)(1-\phi)]$

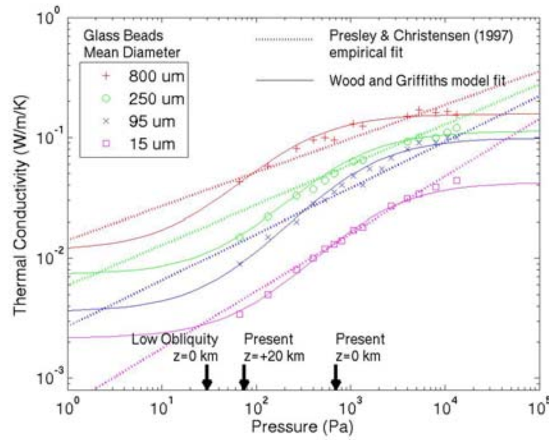

Figure 1: Comparison of empirical fits (dotted lines) and model fits (solid lines) to laboratory measurements (colored symbols) of the thermal conductivity of glass beads as a function of particle size and pore gas pressure [4]. The empirical fits were calculated using the formula derived in [4]: $k = Ap^{0.6}d_{part}^{\{-0.11*\log10(p/B)\}}$, A=7.96E-5, B=1.08E7. The parameter values used for our model fits are given in **Table 1**. Arrows on the x-axis indicate present-day surface pressures at elevations corresponding to the global mean and the top of Olympus Mons, as well as the surface pressure minimum expected at low obliquity.

Table 1

Glass Bead Properties [4]			Model Fit Parameters		
d_{part} (µm)	density (kg/m^3)	porosity* (if ρ_0 = 2550)	porosity	f_{cont}	X
800	2000 ± 100	0.18 - 0.25	0.22	.985	40
250	2000 ± 100	0.18 - 0.25	0.30	.990	40
95	1700 ± 100	0.29 - 0.37	0.33	.997	20
15	900 ± 100	0.61 - 0.69	0.63	1.00	4

*This porosity was calculated using the density of typical soda lime glass – the actual value was not reported in [4].

where k_a and k_s are the thermal conductivities of the air in the pore space and the solid particle material, respectively, and ϕ is the porosity [5]. Comparisons with measurements of the thermal conductivity of terrestrial sands and sandstones with various porosities show that loose, unconsolidated materials tend to have values close to k_{min} whereas more consolidated, rocky materials have values between k_{min} and k_{max}. We have parameterized the relationship between k_{eff}, k_{min}, and k_{max} using a "continuity factor", f_{cont}, such that

(3) $k_{eff} = k_{max} - f_{cont}(k_{max} - k_{min})$, $0 < f_{cont} < 1$.

Heat transfer in a porous medium also occurs by thermal radiation, so that the total thermal conductivity

(4) $k_{tot} = k_{eff} + k_{rad} = k_{eff} + BT^3$

where B = 8E-11 W/m/K^4 [6].

The thermal conductivity of the air in the pore space, k_a, starts to decrease with pressure when the mean free path of air molecules, λ, approaches the average pore size, d_{pore}. To model this effect, we first estimate d_{pore} from the assumed particle diameter and porosity,

(5) $d_{pore} = d_{part} / X$

then calculate the Knudsen number ($Kn = \lambda/d_{pore}$) and use

(6) $k_a = k_{kin}/(Kn+1)$

where k_{kin} is the thermal conductivity of the gas calculated from kinetic theory [7]. So for a given regolith composition, porosity and temperature, there are two free parameters in our model for calculating k_{tot}: f_{cont} and X.

Testing the Conductivity Model. To test this model, we compared it with measurements by Presley and Christensen (1997) [4] of the thermal conductivity of glass beads in CO_2 gas over a range of pressures and particle sizes. As shown in **Figure 1**, we were able to fit the data very closely using the parameters listed in **Table 1** (and $k_s = k_{glass} = 0.937$ W/m/K). Also note that the shape of our model-calculated curves matches the data points more closely than the log-linear empirical fit used in [4]. Although the density of the glass bead samples was measured, their porosity (ϕ) was not, so we treated ϕ as a pseudo-free parameter, constrained by estimates based on a typical solid glass density value. The best-fit values of f_{cont} are all very close to unity as expected for an unconsolidated sample of glass beads. The best-fit values of X ranged from 4 - 40, which can be compared to the value of 6 expected for randomly packed perfect spheres ($\phi \approx 0.42$). Roughness should decrease X, while denser packings should increase X.

1-D Thermal Model: We have implemented the thermal conductivity model described above in a 1-D finite-difference thermal model for the time evolution of Mars' subsurface temperatures over the past 10^6 years to study the effects of varying obliquity and atmospheric pressure (**Figure 2A & 2B**). In this model we define a number of subsurface layers with different thermal and physical properties (**Figure 2D**, for example), representing each layer by 20-40 points

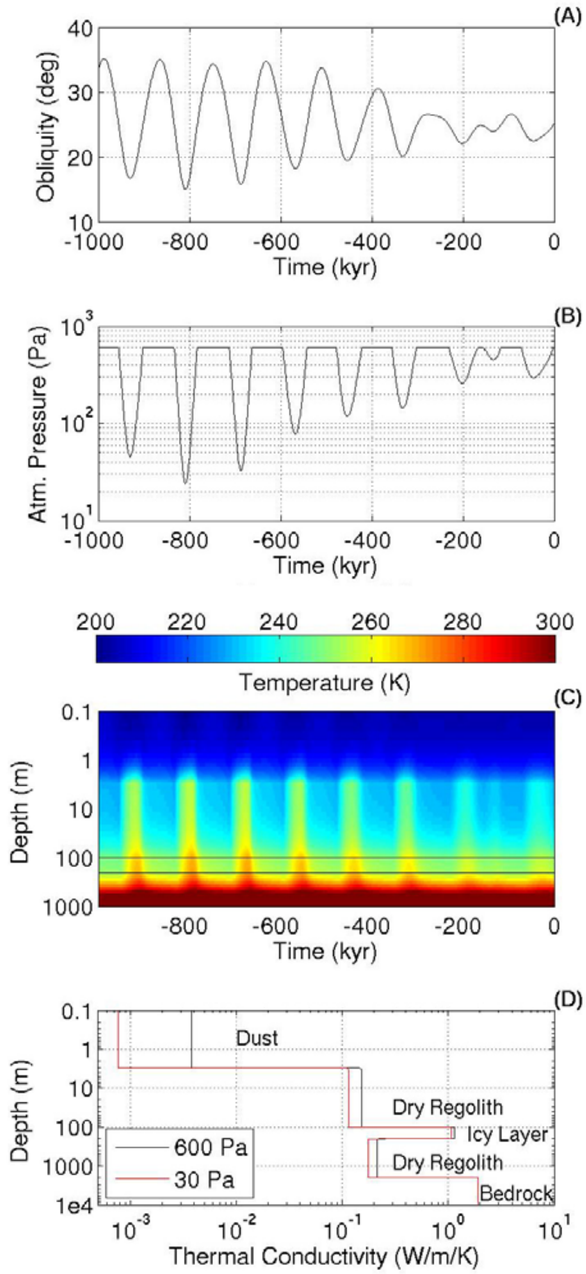

Figure 2: (A) Variation in the tilt angle of Mars' spin axis (obliquity) over the past 1 million years [1]. **(B)** Variation in Mars' atmospheric surface pressure due to the formation of perennial CO_2 ice polar deposits during periods of low obliquity [2]. **(C)** Model-calculated subsurface temperature variations at 45°N due to the reduction in thermal conductivity of porous regolith at low pressures. The surface albedo was 0.25, and the geothermal heat flux was 30 mW m^{-2}. **(D)** Subsurface profiles of thermal conductivity at present-day atmospheric pressure (black line) and at low pressure (red line). [See **Table 2** for the assumed values of particle size and porosity of each subsurface layer.]

on a Chebyshev grid for accurate resolution of the surface and layer interfaces. Time-stepping of the heat diffusion equation is performed to second-order accuracy, including a consistent treatment of the nonlinear surface heat balance, and a semi-implicit scheme is used to ensure numerical stability at large time-steps.

Results for one set of calculations at 45°N latitude are shown in **Figure 2C**. The subsurface structure for this case is described in **Table 2**. At this latitude, near-surface ground ice is not expected to be thermally stable [8,9], and is not observed in MGS GRS data [10]. However, studies of the onset diameter of rampart craters as a function of latitude suggest an ice rich layer may be present at depths of ~100 m [11,12], and models of buried ground ice sublimation rates indicate it could survive billions of years at this depth and latitude [13]. Our calculations show that the temperature of an ice-layer at this depth could increase from ~250 K at high obliquity to nearly 273 K, the melting point of pure water ice, at low obliquity. The presence of salts in the regolith could allow melting at lower temperatures.

We will present additional results for other latitudes and different thicknesses of the surface dust or sand layers to examine the potential effects of this mechanism on the evolution and distribution of water on Mars.

Table 2: Subsurface Layer Structure for Figure 2

Layer Depth Range	**Particle Size (μm)**	**Por-osity**	**Ice Volume Fraction**	f_{cont}
0 - 3 m	2	40 %	0 %	1.00
3 - 100 m	1000	25 %	0 %	0.95
100 - 200 m	1000	25 %	20 %	0.50
200 m - 2 km	1000	25 %	0 %	0.90
2 km - 10 km	Bedrock: k_{th} = 2 W/m/K			

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