

**THE EFFECTS OF OBLIQUITY ON GEOTHERMAL GRADIENTS AND METHANE HYDRATE STABILITY ZONES ON MARS.** M. J. Root<sup>1</sup> and M. E. Elwood Madden<sup>2</sup>, <sup>1,2</sup>University of Oklahoma, School of Geology and Geophysics, 100 E. Boyd St., Suite 710, Norman, OK 73019, <sup>1</sup>mjroot@ou.edu, <sup>2</sup>melwood@ou.edu.

**Introduction:** Methane has been detected in the Martian atmosphere [1-2]. It has been theorized that methane hydrate dissociation is a possible source for this atmospheric methane. Gas hydrates, gas molecules surrounded by a cage of ice, form under high pressures and low temperatures. The feasibility of the presence methane hydrate reservoirs of the requisite size and depth to account for the observed methane release was investigated by Root et al. [3]. This work showed that methane hydrates could exist at a latitude coinciding with the observed methane plumes and at a depth allowing for dissociation and diffusion to occur, especially if the methane was mixed with H<sub>2</sub>S.

This work builds upon the previous model to show how changes in temperature due to obliquity affect depths to the hydrate stability zone (HSZ). Results of this study will be used in the future to examine the kinetics of hydrate dissociation and diffusion within this system.

**Current Methane Hydrate Depth:** Present surface temperatures as well as modeled surface temperatures for varying obliquities ([4], [5], [6]) are required to determine the depth to the HSZ. Temperatures were compiled for the equator, ±30°, ±60°, and ±90° latitude.

Pressure and geothermal gradients were calculated using

$$P = P_0 + \rho g d; T = T_{0, \text{lat}} + \frac{h}{k} d$$

where  $P_0$  is the average surface pressure of Mars (636 Pa),  $\rho$  is density,  $g$  is the Martian gravity (3.7 m/s<sup>2</sup>) and  $d$  is depth in m;  $T_{0, \text{lat}}$  is the surface temperature at the latitude of interest,  $h$  is the geothermal heat flow (0.03 W/m<sup>2</sup>), and  $k$  is the thermal conductivity of the ice/hydrate cemented regolith [7].

Thermal conductivity and density both depend on the composition of the ground, which is controlled by porosity. We assumed the rock was basalt and that the pore space was filled with ice near the surface, then hydrate, then finally liquid water at depths with high enough temperature and pressure conditions. Changing pore space with depth due to compression was calculated using

$$p = p_0 e^{-\frac{d}{K}}$$

where the porosity ( $p$ ) is dependant on  $p_0$ , the initial porosity (assumed 35%), depth ( $d$ ) in km, and  $K$ , the porosity decay product (2.82 km) [7]. For example, the porosity changes were applied to the density and thermal conductivity to ice using

$$\rho_{ice} = p\rho_{ice} + (1-p)\rho_{basalt}$$

$$k_{ice} = pk_{ice} + (1-p)k_{basalt}$$

where  $\rho_{ice}$  is 920 kg/m<sup>3</sup>, ( $\rho_{hyd}$  is 950 kg/m<sup>3</sup>,  $\rho_{liq}$  is 1000 kg/m<sup>3</sup>),  $\rho_{basalt}$  is 3000 kg/m<sup>3</sup> and  $k_{basalt}$  is 2.0 W/m, ( $k_{ice}$  is 2.8 W/m,  $k_{hyd}$  is 0.45 W/m, and  $k_{liq}$  is 0.58 W/m) [8].

Since the depth to the HSZ is unknown, the geothermal and pressure gradients were calculated iteratively, initially assuming that the pore spaces were filled with liquid water. The thermodynamic model HYDOFF [9] was used to determine the P and T conditions of the methane HSZ. With this estimate, the values for ice and hydrate were used at their expected depths and the geothermal and pressure calculations were re-evaluated. HYDOFF was used to confirm the location of the hydrate stability field. This process was repeated with a resolution of 1/10<sup>th</sup> m to locate the top of the HSZ. The depths for the top of the HSZ for the four latitudes of interest are shown in Figure 1. The depth to the top of the HSZ decreases with increasing latitude. This trend is the result of surface temperature changes with latitude. The equator is warmer than the poles causing a deeper, thinner HSZ at the equator as opposed to a shallow, thick HSZ at the poles.

**Equilibrium Model of the HSZ with Changing Obliquity:** The model was also calculated for obliquities ranging from 10° to 50° based on surface temperatures provided by Fanale et al. [4], Mellon and Phillips [5], and Jakosky et al. [6]. As obliquity increases, the amount of insolation at the equator decreases and increases at the poles, causing temperatures to cool at the equator and warm at the poles. The model results for 30° and 60° do not seem as definitive as the results for 0° and 90°. This may be caused by a smaller change in insolation at these locations due to being located near mid-latitudes. The models show the best agreement for surface temperatures at the poles. Some observable trends include a shallowing of the HSZ at the equator as obliquity increases, while the depth to the top of the HSZ increases at the pole as obliquity increases. There is also a slight deepening of the hydrate at the equator as obliquity increases and a shallowing of the HSZ at the poles with increasing obliquity.

**Thermal Gradient Kinetics:** Temperature changes due to obliquity are not instantaneous at depth. Heat must be transferred through the subsurface. In order to model the time it takes for this temperature adjustment to occur and how the geothermal gradient changes over time with obliquity, Lachenbruch and Marshall's [10] model was adapted for use under Martian conditions. The temperature profile as a result of a

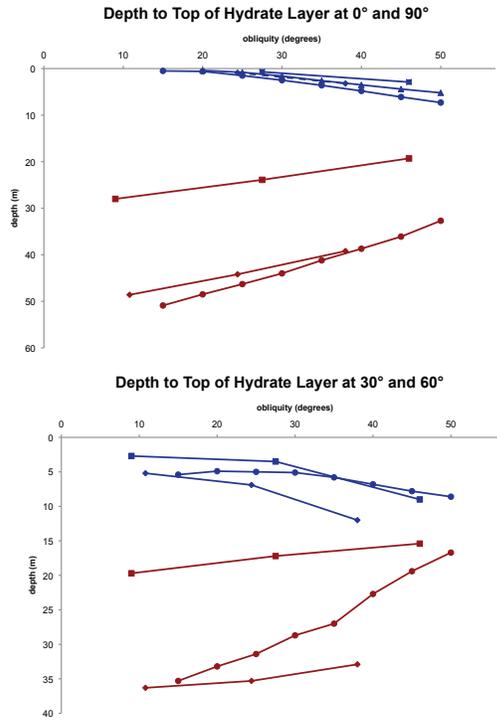


Figure 1. Depth to the top of the methane HSZ for 0° (red), and 90° (blue) on the left and 30° (red) and 60° (blue) on the right for surface temperatures from [4] (current obliquity - diamonds, pre Tharsis obliquity - squares), [5] (circles) and [6] (triangles, 90° only).

change of surface temperature is given by

$$\theta = \theta_0 + \Gamma_0 z + T(z,t)$$

where  $\theta$  is temperature,  $\theta_0$  is the initial surface temperature,  $\Gamma_0$  is the original thermal gradient,  $z$  is depth, and  $T(z,t)$  is the temperature anomaly with depth and time. The temperature anomaly can be calculated with the equation

$$T(z) = D\Gamma\left(\frac{1}{2}n + 1\right)2^n i^n \operatorname{erfc} \frac{z}{\sqrt{4\alpha t^*}}$$

where  $D$  is the change in surface temperature,  $\Gamma$  indicated the gamma function,  $n$  is a number greater than 0 describing the shape of the geothermal gradient (a value of 2 was used as the temperature trend between 30° to 15° obliquity is almost linear),  $\alpha$  is the thermal diffusivity, calculated by dividing the thermal conductivity by the density multiplied by the specific heat,  $t^*$  is the time over which the temperature changed ( $3.69 \times 10^4$  yrs), and  $i^n \operatorname{erfc}$  is the  $n$ th integral of the error function.

By varying  $t^*$ , the time delay in adjusting the geothermal gradient can be visualized (Figure 2).

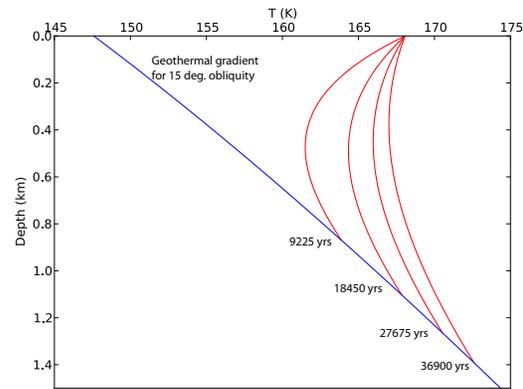


Figure 2. A comparison of the thermal gradient at 15° obliquity (blue) and thermal gradients resulting from a temperature change over varying times.

The longer the system takes to reach the new temperature, the deeper the thermal gradient is altered from the original thermal gradient. If the rate of hydrate dissociation and diffusion is greater than the rate of thermal diffusivity (values used in model), the HSZ will adjust with the geothermal gradient. If the thermal diffusivity is faster than the rate of hydrate re-equilibration, metastable may be preserved in the HSZ.

**Conclusion:** As obliquity changes, the surface temperature and therefore the thermal gradients also change. If these thermal gradient changes reach the hydrate stability field, the methane hydrate could dissociate and methane will be released into the atmosphere. Hydrates located at the poles would be more susceptible to changes in the thermal gradient as they are located closer to the surface.

**References:** [1] Mumma, M. J. et al. (2009) *Science*, 323, 1041-1-45. [2] Formisano, V. et al. (2004) *Science*, 306, 1758-1761. [3] Root, M. J. et al. (2010) *LPS XXXXI*, Abstract #1705. [4] Fanale, F. P. (1986) *Icarus*, 67, 1-18. [5] Mellon, M. T. and Phillips, R. J. (2001) *JGR*, 106, 1-15. [6] Jakosky, B. M. (1995) *JGR*, 100, 1579-1584. [7] Clifford, S. M. (1993) *JGR*, 98, 10,973-11,016. [8] Mellon, M. T. (1996) *Icarus*, 124, 268-279. [9] Sloan, E. D., Koh, C. (2008) *Clathrate Hydrates of Natural Gases*, CRC Press, New York. [10] Lachenbruch, A. H. and Marshall, B. V. (1986) *Science*, 234, 689-696.