

**A COUPLED SUBSURFACE ATMOSPHERE BOUNDARY LAYER MODEL OF H<sub>2</sub>O ON MARS**, A. P. Zent, SETI Institute and NASA Ames Research Center, R. H. Haberle, NASA Ames Research Center, Moffett Field, Ca 94035, H. Houben, Space Physics Research Institute, Sunnyvale, Ca. 94087, B. M. Jakosky, LASP, University of Colorado, Boulder, Co. 80309.

We have developed a coupled subsurface-atmosphere boundary layer model of H<sub>2</sub>O exchange on Mars. Our objectives in constructing this model are: a) to identify the physical processes that control exchange of water between the atmosphere and regolith over a diurnal timescale on Mars, and to define physical parameters that can be used predictively; b) to understand the intensity as well as temporal and spatial variability of the H<sub>2</sub>O flux through the martian surface; c) to develop a description of how exchange patterns vary throughout reasonable parametric space; and d) to advance our understanding of the martian annual H<sub>2</sub>O cycle.

**Model:** The atmospheric boundary layer model is taken from Haberle *et al.* 1990, and is described in detail by those authors. It is a radiative-convective model which allows for the radiative effects of dust. The model is similar to that by Flasar and Goody (1976), however the model has an improved treatment of turbulence, and provides a self-consistent calculation of surface stresses, which help regulate the exchange of water between the surface and atmosphere.

The subsurface model consists of a thermal component and a H<sub>2</sub>O transport component. The heat conduction equation was solved in order to find subsurface temperatures. Viking data was used to constrain the thermal constants. Boundary conditions are found from the atmosphere model, which calculates the radiative and sensible heat fluxes. The surface temperature is then calculated assuming that the radiative, sensible, conductive, and latent heat fluxes sum to zero. The lower boundary condition is found by extending the solution region obtained to depths greater than the penetration depth of the annual thermal wave. The thermal gradient at the lower boundary is then assumed to be the quotient of the heat flux, calculated assuming a chondritic Mars, and the conductivity.

There are four assumptions that underlie the model of H<sub>2</sub>O transport through the regolith. The first is that vertical transport of H<sub>2</sub>O through the regolith is according to Fick's law. We further assume that water in the regolith must exist in one of three discrete states, either as vapor, adsorbate, or ice.

$$\sigma = f\gamma + \alpha(\gamma, T) + \iota \quad (1)$$

Where  $\sigma$  is the total density of water per cubic meter of regolith,  $f$  is the porosity of the regolith,  $\gamma$  is the vapor density per cubic meter of gas,  $\alpha$  is the adsorbed phase, and  $\iota$  is the ice phase.

Some assumption must be made regarding the form of the expression  $\alpha(\gamma, T)$ . Our third assumption is that

$$\alpha(\gamma, T) = \frac{\beta P^{0.51}}{\exp(\delta/T)} \quad (2)$$

Where  $\beta = 2.043 \times 10^{-8}$ , and  $\delta$  is -2679.8 (Fanale and Cannon, 1971).  $P$  is the partial pressure of H<sub>2</sub>O, leading to our final assumption, that the ideal gas law is obeyed by H<sub>2</sub>O.

The evolution of the population in a finite volume of regolith is controlled by

$$\frac{d\sigma}{dt} = \frac{d\phi}{dz} = \frac{\partial}{\partial z} D \frac{\partial \gamma}{\partial z} \quad (3)$$

Where  $\sigma$  is the total concentration of H<sub>2</sub>O in the regolith,  $\gamma$  is the vapor phase concentration, and  $\phi$  is the flux. The upper boundary condition is a flux, which is calculated by assuming that the vapor pressure of H<sub>2</sub>O at the mathematical surface is controlled by adsorption. That permits the atmosphere layer model to calculate a gradient and resulting flux through the surface via stability functions from Monin-Obukhov similarity theory. The lower boundary condition for H<sub>2</sub>O is zero flux.

The model is initialized with no H<sub>2</sub>O in the atmosphere and 2.1 kg m<sup>-3</sup> throughout the regolith. We carry out the calculation until the day to day variation of H<sub>2</sub>O in the atmosphere column is less than 0.01%. Accordingly, the atmospheric H<sub>2</sub>O column is stable over a timescale longer than the diurnal radiation pattern remains constant. We note that equilibration typically

requires 12 to 18 sols, providing the first estimate of the timescale for the atmosphere and regolith to equilibrate.

**Results:** As our baseline case, we take the latitude and season that the Viking One lander touched down; albedos and thermal inertias are consistent with the Viking data.

Beginning at midnight, the atmosphere is quite stable, the only turbulence is due to wind shear (Fig. 1). The regolith cools progressively throughout the night, increasing its adsorptive capacity (Eq. 2 and Fig. 2). The water vapor mixing ratio in the lower part of the atmosphere decreases as the night progresses because the cooling regolith is scavenging H<sub>2</sub>O; due to the stability of the atmosphere, supply of water to the regolith is effectively diffusion limited.

Sunrise occurs just before 0600 hours, and the soil begins to warm. Although the atmosphere is not yet convective, there is a thin layer of atmospheric mixing that brings additional water down to the surface, where the warming regolith continues to adsorb more water. Although physical adsorption is inhibited at higher temperatures, we predict continued adsorption after sunrise because the regolith is not in equilibrium with the atmosphere as a whole, but only with the strongly depleted lowest tens of meters.

Even after the atmosphere becomes fully convective, about 0800 hours, there is still a short period where the atmosphere is supplying water to the regolith. The flux *into* the soil increases dramatically when the atmosphere begins to convect freely (Fig. 3). It isn't until about 1000 hours that the increasing temperature and adsorbed water content raise the pore pressure of water to atmospheric levels and the flux reverses.

During the day the atmosphere is fully convective, and there is only a very shallow vertical gradient in atmospheric water. The regolith communicates during the day with the entire boundary layer, and approximate equilibrium exists. Even before the surface temperature peaks, the flux to the atmosphere begins to drop as the amount of near-surface H<sub>2</sub>O is depleted. Again in the late afternoon, there is a strong pulse of H<sub>2</sub>O *into* the regolith, as the atmosphere is still convecting vigorously, but the regolith is now cooling rapidly. Finally, once the surface cools below the temperature of the atmosphere, about 1650 hours, convection stops, the atmosphere stabilizes, and the slow scavenging of H<sub>2</sub>O from the lowest tens of meters of the atmosphere begins again.

We have examined the effects of albedo, thermal inertia, latitude, atmospheric optical depth, and pore size on our baseline model, and will report those results as time permits.

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Flasar, F. M. and R. M. Goody, 1976, *Planet Space Sci.* **24**, 161-181.

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Subsurface Adsorbed Water  
Baseline Case

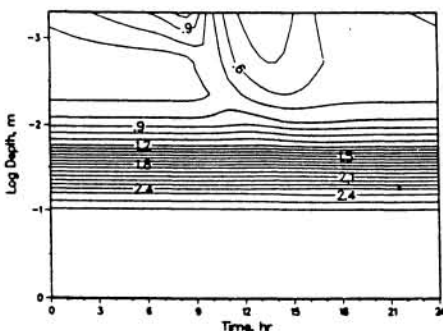


Figure 2. Subsurface adsorbed water profile vs. time.

Water Vapor Mixing Ratio  
Baseline Case

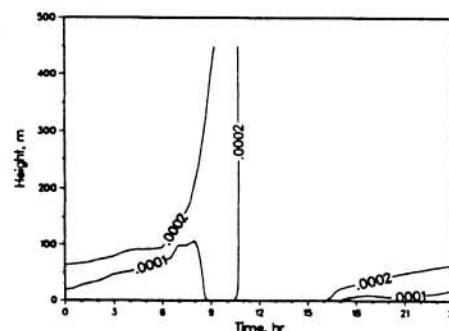


Figure 1. Atmospheric water vapor mixing ratio vs. time.

Water Vapor Flux  
Baseline Case

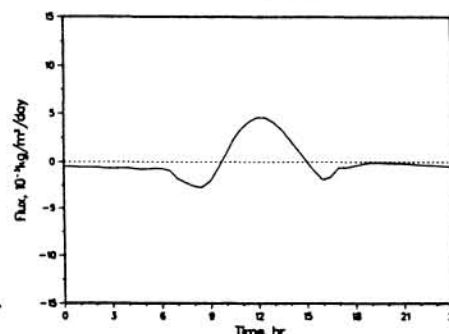


Figure 3. H<sub>2</sub>O flux vs. time. Flux greater than zero is up.

INVESTIGATION OF PLESSITE IN IRON METEORITES AND LABORATORY Fe-Ni(P) ALLOYS; J. Zhang, D.B. Williams, and J.I. Goldstein, Dept. of Materials Science & Engineering, Lehigh University, Bethlehem PA 18015, USA

Plessite in iron meteorites is a two phase structure with an fcc precipitate phase in a bcc matrix.<sup>1,2</sup> After Fe-Ni martensite forms during slow cooling, the martensite decomposition occurs at different temperatures. The morphology of the precipitates and the Ni content of both precipitate and matrix vary with the local average Ni composition of the plessite.

In this study, the plessite structure of two octahedrites, Carlton and Grant, was characterized using the analytical electron microscope (AEM). The composition of the taenite precipitates in various regions of plessite which have 9 to 13 wt% and 15 to 20 wt% Ni composition were measured using an x-ray energy dispersive spectrometer (EDS) in the AEM. To understand the phase transformation processes which occurred during the plessite formation, an experimental set of Fe-Ni binary and Fe-Ni-P ternary alloys were made and analyzed also using the AEM. The alloys, which have 15 to 30 wt% Ni (0.2 -0.3 wt% P for ternary alloys), were first homogenized at 1200°C and quenched to liquid nitrogen temperature to form martensite. They were then isothermally heat treated for 60 to 400 days in the temperature range from 450°C to 300°C. Two phase structures, which are similar to those of plessite, were formed in these alloys.

Two types of taenite precipitates are observed in the plessite region. The first type of precipitates is primarily formed at plessite regions of low Ni content (9-13 wt%). They are formed at the original martensite grain boundaries and have a relatively large size (50-200 nm wide). The precipitates in the low Ni region have a ~50 to 54 wt% Ni composition and are in equilibrium with a matrix phase of ~4 to 5 wt% Ni. The second type of precipitates is mostly formed in plessite regions of higher Ni concentration (15-20 wt%). They are formed inside the original martensite grains along all crystallographically equivalent directions with the precipitate {111} plane parallel to the matrix {110} plane. Their sizes are very small, typically 10-20 nm wide. The precipitates in the high Ni region have ~57 to 60 wt% Ni and are in equilibrium with a matrix phase of ~10 wt% Ni. Fig.1 shows an EDS Ni composition profile of a precipitate taken in an 18 wt% Ni region of the plessite of Carlton. Precipitate phases in both plessite regions in Carlton have the ordered fcc structure of Tetrataenite. However, precipitate phases in both plessite regions in Grant have a disordered fcc structure. This lack of the Tetrataenite phase in Grant is most likely due to the shock effect.

The microstructures of laboratory alloys have a similar morphology to that of the plessite regions of the same average Ni composition. Fig.2 shows TEM images of a 25 wt% Ni plessite region of the Carlton meteorite and a 30 wt%Ni alloy heat treated at 300°C for 370 days respectively. The precipitates are the dark platelets in the TEM images. The precipitates in the alloy have an fcc structure and no ordering is observed. This indicates that the ordering reaction is a much slower process than the martensite decomposition or the ordering temperature of FeNi is lower than 300°C rather than the reported 320°C. The compositions of the precipitate and the matrix of the alloy, measured using the EDS in the AEM, are ~57 wt% and ~21 wt% respectively. The microstructure of the alloy of 30 wt% Ni heat treated at 400°C is the same as that shown in Fig. 2 except that the sizes of the precipitates are larger (>50 nm wide) due to the higher diffusivity at the higher temperatures. The Ni composition of the precipitate and matrix in the 400°C alloy is ~50 wt% and ~8 wt% respectively.

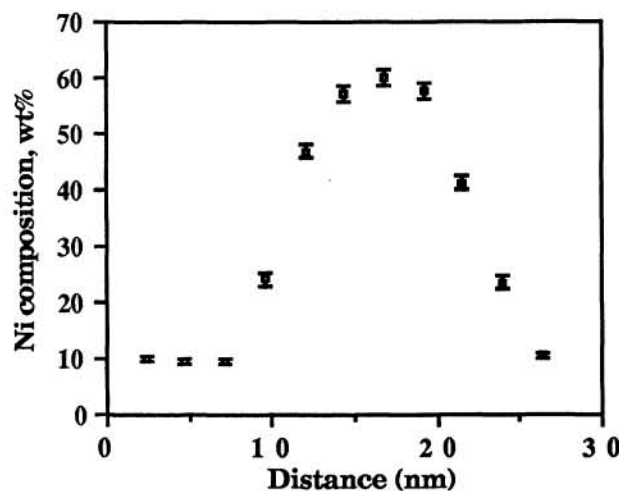
The Fe-Ni phase equilibria measured in the decomposed martensite alloys can be used to explain the difference in Ni composition between precipitates in the high Ni and low Ni plessite regions. In the slow cooling process, the martensite is formed, and therefore,



decomposed at relatively higher temperatures for the low Ni plessite region. In this region, the Ni composition of the precipitates increases gradually to ~50 wt% as cooling continues. The FeNi ordering transformation then occurs and stabilizes the precipitate composition at ~52 wt% Ni. In contrast, the martensite transformation occurs at lower temperatures in the high Ni plessite region. The precipitates formed by martensite decomposition at low temperatures have a Ni composition higher than 52 wt%. The FeNi ordering occurs after the precipitation and, therefore, the precipitates containing Tetrataenite have a Ni composition of 57 to 60 wt%.

The measured matrix composition in the low Ni plessite region is consistent with the current Fe-Ni phase diagram<sup>3</sup>, however, that of the high Ni region is not. If we assume that the meteorites have enough time to reach equilibrium, the higher matrix Ni composition in the high Ni plessite region probably arises because of the effect of very sharp edge of the precipitate (<10 nm radius) on the local equilibrium composition, which termed Gibbs-Thomson effect<sup>4</sup>. The matrix Ni composition of the alloys is a complex function of the heat treatment. All compositions are higher than the predicted values in the phase diagram. In general, the lower the temperature and the finer the precipitates, the higher is the matrix composition. No significant composition gradient is detected to a spatial resolution of 2 nm. This indicates that the martensite decomposition process in Fe-Ni at low temperatures is interface reaction controlled. The much higher Ni composition of the matrix phase of the laboratory alloys is clear evidence that the heat treatment time is not long enough to bring the alloys to fully equilibrium.

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Fig.1 EDS Ni composition profile of a precipitate in an 18% Ni area of meteorite Carlton. Fig.2 TEM images of a) a plessite region of Carlton meteorite, b) Fe-30wt%Ni alloy, 300°C for 370 days.