The distribution and behavior of Martian ground ice during past and present epochs

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Abstract. Mars undergoes significant oscillations in its orbit, which will have a pronounced effect on its climate and, in particular, on the behavior of subsurface water ice. We explore and map the behavior of ice in the Martian near-surface regolith over the past 1 m.y. using a diffusion and condensation model presented in an earlier paper, with two primary modifications to include orbitally induced variations in insolation and atmospheric water abundance. We find that the past behavior of ground ice differs significantly from that at the present epoch, primarily the result of high-amplitude oscillations in obliquity (presently 25°). In midlatitude and equatorial regions, ground ice will condense from atmospheric water during times of higher obliquity, filling the top few meters of the regolith with significant amounts of ice. At an obliquity of 32°, ground ice becomes stable globally. During times of lower obliquity, ground ice will sublime and diffuse back into the atmosphere, desiccating the regolith to a depth of about 1 to 2 m equatorward of 60° to 70° latitude. In the high-latitude regions these oscillations are considerably subdued. Below this depth of cyclic saturation and desiccation a long-term stability of ice exists in some geographic regions. We present a map of the distribution of ice expected at the present epoch. Cyclic exchange of water between the global regolith and polar regions will have significant implications for surface geology and the polar layered deposits.

Introduction

Changes in climate on Earth, such as ice ages, are known to be influenced by oscillations in its orbit [e.g., Imbrie and Imbrie, 1980], which come about by gravitational interactions with other planets and the Sun. The amplitudes of these oscillations are small relative to those that occur on Mars. Therefore, it is possible that orbitally induced changes in the Martian climate are considerably more impressive. In an earlier paper [Mellon and Jakosky, 1993] (hereafter Paper 1), we showed that at high obliquities the combination of increased sublimation of water from the pole and colder equatorial and midlatitude temperatures allows ground ice to be stable globally. Conversely, at low obliquities the reduction in atmospheric water and warming of equatorial and midlatitude regions will restrict ice stability to the polar regions. Questions remain as to whether diffusion of atmospheric water vapor is capable of populating the near-surface regolith with ice before the orbit and climate change toward more unfavorable conditions and how the orbital history will impact the present distribution of ice.

All the Martian orbital elements oscillate with characteristic timescales. For our interests these can be reduced to three fundamental parameters. The obliquity, Θ (tilt of the spin axis relative to the orbital plane) and the eccentricity (the degree of ellipticity of the orbit) experience similar oscillations with timescales around 10⁵ years. The L₅ of perihelion (the arcocentric longitude of the Sun, or season, at which perihelion occurs) processes with periods between 50,000 and 53,000 years. Other periodicities occur to modulate the amplitudes and cause small-scale changes (see Figure 1). In Paper I we showed that the diffusion of water vapor is sufficiently rapid to allow significant quantities of atmospheric water to be transported into the regolith and to condense as ice or to allow existing ice from an earlier epoch (orbital regime) to sublime and diffuse into the atmosphere on these same timescales. Here we examine
the effect of such orbitally induced changes in climate on the thermal and diffusive stability of near-surface ground ice.

There are two primary responses of the Martian climate to orbital variations which will affect the behavior of ice. The first is a temperature change due to the redistribution of insolation with latitude. Oscillations in the obliquity will change the distribution of insolation between equatorial and polar regions. For example, at an obliquity higher than the present value (25°), more solar energy will be deposited in the polar regions raising the annual mean temperature, while less energy reaches the midlatitude and equatorial regions resulting in lower annual mean temperatures. The situation is reversed during times of lower obliquity. The obliquity dominates the overall changes in temperature. Precession of the Ls of perihelion and oscillations in the eccentricity change the seasonal distribution of solar energy between the northern and southern hemispheres. Although these parameters are less important, they nonetheless have an effect and will be discussed later.

Second, as a by-product of increasing the mean polar temperature during higher obliquities and decreasing it during lower obliquities, temperature changes will affect the summertime sublimation of the residual water-ice polar cap [Toon et al., 1980, Jakosky et al., 1993, 1995]. An increase in either the polar surface temperature or the length of the summer season will allow more water to sublime into the atmosphere, thus increasing the atmospheric water abundance and affecting the rate and direction of diffusive transport of water vapor in exchange with the regolith.

Previously, Fanale et al. [1986] examined the long-term behavior of ground ice over Martian geologic history. Their results suggested that such changes in atmospheric water vapor abundance can be important in calculating the long-term stability of ice in the Martian regolith. They also indicated that for an initially saturated regolith, ice poleward of about 40° latitude would remain stable throughout Martian history, while ice equatorward of this latitude would sublime and be lost to the atmosphere, consistent with earlier models (see Paper I and references therein). However, their model only allowed for the sublimation and condensation of ice at the top of the subsurface boundary between an ice-saturated regolith and an ice-free regolith, rather than allowing the vapor/ice equilibrium conditions to dictate the behavior of ice at all depths simultaneously. Therefore, their results did not indicate how ice would behave in the near surface (top several meters) during epochs favorable to ice condensation from atmospheric water. However, in a similar model, Zent et al. [1986] included the effects of vapor/ice equilibration at all depths, while examining the seasonal and climatological behavior of water at a few latitudes in the high-latitude regolith. Their results suggested that some water would exchange between the regolith and the atmosphere, but they did not consider the equatorial regolith or the range and geographic distribution of the regolith's thermal properties.

Here, we explore the behavior of near-surface ground ice on orbital and geologic timescales using the thermal and molecular diffusion models described in Paper I. The models allow for condensation at any depth where the equilibrium conditions are met and include the geographic distribution of thermal properties, enabling us to map the behavior of ground ice. We will show that atmospheric water will diffuse into and condense within the near-surface regolith, globally, during periods of high obliquity, while during periods of low obliquity the top 1 to 2 m of the regolith will become desiccated equatorward of 60° to 70° latitude. We will also show that, below this “active layer”, can exist a region of relative long-term stability, consistent with the earlier models. The model presented in this paper differs from the model of Fanale et al. [1986] by (1) using the latest orbital history model, (2) using an improved model of atmospheric water abundance (based on polar cap sublimation rates), (3) allowing ice to condense at any depth where the equilibrium conditions are met, (4) including diffusion through a regolith partially choked by pore ice, and (5) by considering a different set of thermal and diffusive parameters. In addition, we map geographic variations in the behavior of ground ice caused by the geographic variations in the regolith thermophysical properties.

Mars Orbital History

As input to our thermal and molecular diffusion models, we use the orbital history of Mars. Specifically, this input consists of a linear perturbation model [Ward, 1974; Bills, 1990] with secular Fourier components from Laskar [1988]. From this model we calculate the history of obliquity, eccentricity, and the Ls of perihelion. Figure 1 shows the orbital history used for the past 2.5 m.y. (about 1.3 million Martian years). Obliquity

![Figure 1. The orbital history of Mars for the past 2.5 million Earth years (about 1.3 Martian years). Currently, the obliquity is 25.19°, eccentricity is 0.0934, and the Ls of perihelion occurs at 251°.](image-url)
can be seen to vary between about 15° and 35°, with a current value of about 25°; while most of this history is marked by a high-amplitude oscillation, the past 300 kyr have been relatively quiescent. These orbital parameters are used, for each epoch (1 or 5 kyr increment), to compute the insolation at any time of day or year.

There are several areas of uncertainty in this orbital history. The linear perturbation model is considered less accurate than a more complex numeric integration model; however, this uncertainty increases as we go back in time, leaving the most recent history well represented [Bills, 1990]. For our purposes, the past 1 m.y. in Figure 1 is adequate, and for models examining a longer 2.5 m.y. history, the exact phase and amplitude of the oscillations are unimportant. In addition, any orbital model is limited to the accuracy of the secular Fourier components of Laskar [1988]. To compute the obliquity, we use an axial precession rate, α, of 8 arcsec/yr. The uncertainty in the axial precession rate reflects our lack of knowledge about the Martian moment of inertia; estimates of α range from 8 to 12 arcsec/yr. While the longer-term obliquity history is very sensitive to the choice of α (allowing for excursions in obliquity between about 0° and 51°), the more recent past is relatively insensitive to our choice of α [Bills, 1990]. Furthermore, the orbits of all the inner planets were found to be chaotic on timescales of 10⁷ years [Laskar, 1989], and specifically the obliquity of Mars was found to behave chaotically on these timescales and may have experienced excursions between 0° and 60° [Laskar and Robutel, 1993; Touma and Wisdom, 1993]. The chaotic nature does not begin to become important until 10 m.y. in the past and therefore should not affect our results; however, it is interesting to note the range of possible obliquities when considering the results presented here. In addition, the results of our numerical models (discussed below) for the past million years or so should be considered representative of the style of ground-ice behavior throughout much of Martian history.

**Thermal and Molecular Diffusion**

To determine the behavior of water in the near-surface regolith, we employ the thermal and molecular diffusion models described in detail in Paper I. Briefly, the thermal behavior is found by a standard time-marching finite difference solution to the diffusion equation with the appropriate boundary conditions. The primary components are seasonal CO₂ surface frost, solar heating, and seasonal subsurface heat storage. Geographic variations in the thermal behavior are controlled by changes in latitude and by the geographic distribution of thermal inertia and thermally derived albedo from Palluconi and Kieffer [1981]. The geothermal gradient was shown in Paper I to be largely unimportant for near-surface ice stability and is ignored.

The molecular diffusion of water vapor is found by a similar solution to the vapor-diffusion equation with the boundary conditions controlled by the atmospheric water at the surface and a nonporous layer at some small depth below the surface. The phase partitioning between ice, vapor, and adsorbate on a basalt powder regolith is included. Models are run at specific geographic locations for which the surface thermal properties have been derived by Palluconi and Kieffer [1981]. In each model run the regolith begins completely devoid of water. The regolith is assumed to have a porosity of 40%, allowing a maximum of 0.37 g/cm³ of ice to accumulate. The surface and subsurface temperatures from the thermal diffusion model are used as input to the molecular diffusion model. Feedback between ice concentration and the thermal diffusivity is not included. Paige [1992] showed that the effect of adding high-conductivity ice to the regolith at stable depths is to move the top of the stable region closer to the surface. This behavior is explained by the high-conductivity ice distributing heat over a larger subsurface depth, effectively reducing the maximum subsurface temperatures. For this reason the results presented here might slightly overestimate the depth of ice stability; otherwise, both qualitative and quantitative results remain unaffected. We choose to ignore this effect for reasons of numerical tractability. In addition, we ignore variations in the ambient atmospheric pressure, which may vary with obliquity [Toon et al., 1980], since it will only have a second-order effect on the diffusion coefficient, particularly in the presence of pore ice (see Paper I).

The atmospheric water-vapor density proves to be an important boundary condition for the molecular diffusion model. While the seasonal behavior of the atmospheric water is likely to be a combination of interactions between the atmosphere and nonatmospheric reservoirs (seasonal and residual polar ice and regolith ice and adsorbate) [Jakosky, 1983; Haberle and Jakosky, 1990], the annually averaged (or global) behavior appears to be controlled by interactions with the polar caps [Jakosky, 1985; Jakosky and Haberle, 1992]. At the current epoch seasonal interactions consist of (1) northern summer sublimation of exposed residual water ice and of ice cold-trapped within seasonal CO₂ frost, (2) meridional transport of water away from the polar regions (although the quantity and distance transported during one season may not be large [Haberle and Jakosky, 1990]), and (3) possible return of water to the northern cap during other seasons [e.g., Burns, 1990]. Averaged over a year, the net result of the seasonal behavior is a fairly smooth gradient in atmospheric water vapor (as seen by Viking) from north to south and a global mean around 10 pt μm (precipitable micrometers) [Jakosky and Farmer, 1982; Jakosky and Barker, 1984]. This gradient suggests that the north residual polar cap is a net source of water, with the south cap being a net sink [Jakosky and Farmer, 1982; Jakosky and Haberle, 1992]. Although our understanding of the seasonal processes is limited at best, whatever sublimation and transport occur at present clearly results in a mean of approximately 10 pt μm of atmospheric water.

At other epochs, changes in the distribution of insolation will change aspects of the seasonal behavior
and thus the annually averaged behavior. Specifically, a change in obliquity will change the sublimation rate of the residual polar ice. For example, at 35° obliquity the net summertime sublimation might be several hundred times higher than at present, while at 15° obliquity the sublimation might be more than 10 times lower than at present [Toon et al., 1980; Jakosky et al., 1993]. The transport of water to and from the polar regions is more uncertain. However, if the sublimation is increased at higher obliquities, the average atmospheric water abundance should also increase. In a single season a factor of 2 increase in sublimation may not greatly affect the annual average atmospheric abundance, since even at the current epoch the timescale for global transport of water vapor is longer than a year [Jakosky and Haberle, 1992]; however, after a few years at the increased level of sublimation, we would expect the atmosphere to increase in water abundance by approximately the same factor. Differences might occur, depending on the degree that nonpolar reservoirs interact with the atmosphere. Certainly, since atmospheric transport is more rapid than orbital evolution, the atmosphere should continually track the changes in the polar insolation.

In the absence of a more detailed understanding of the seasonal and annual behavior of water in the Martian atmosphere at the current epoch as well as other epochs, we choose the simplest approach of assuming that an increase or decrease in the polar cap sublimation rate is matched by an equivalent magnitude increase or decrease in the mean atmospheric water abundance. By assuming that 10 pr μm is representative of the current epoch, obliquity of 25.19°, and scaling the atmospheric water abundance to a fit to sublimation rate at other epochs calculated by Jakosky et al. [1993], we obtain the atmospheric column abundance,

$$n_{atm} = D e^{AX^2} B X^C$$

in pr μm, where $X = \ln(\Theta)$. For obliquities lower than the present, $A = -1.83, B = 16.6, C = -28.7$, and $D = 0.08$ pr μm. For the current obliquity and higher, $A = -15.3, B = 117, C = -212$, and $D = 0.0199$ pr μm. Figure 2 shows the atmospheric water abundance given by equation (1). About a factor of 2 fluctuation in $n_{atm}$ can occur due to variations in the other orbital parameters [see Jakosky et al., 1993]. As these are smaller than other uncertainties in our assumptions, we ignore them.

Since diffusion of water vapor in and out of the regolith depends on the atmospheric vapor density near the surface (see Paper I), we convert the column abundance from equation (1) to a near surface density by assuming that water vapor is distributed uniformly in the atmosphere. This assumption is consistent with observations and cloud microphysics at the present epoch [see Jakosky and Haberle, 1992]. The right side of Figure 2 shows the frost point (the temperature at which the near-surface atmospheric water will saturate). When model surface temperatures at any instant drop below the frost point at a given obliquity, the atmosphere is assumed to saturate, and the vapor density is set to the saturation density for that temperature. This procedure effectively allows a thin surface frost to form during colder seasons or at night.

Many processes may act to reduce or enhance the polar sublimation rate and thereby the global atmospheric water abundance. The polar surface may form a lag deposit (dust and sand left behind on the surface after the removal of volatiles). A few-centimeter-thick lag would act to retard sublimation [Höfstadter and Murray, 1990], while a thin deposit might only lower the albedo, raising surface temperatures and thus enhance sublimation. It is also possible that windblown sand might remove surface dust, cleansing the polar ice (J. M. Moore, personal communication, 1994). Additionally, the formation of clouds and precipitation at other epochs might change the vertical distribution of water, as well as the incident solar energy. Our atmospheric model is necessarily simple, given our lack of knowledge concerning the behavior at the present epoch and the increased uncertainty extending it to other epochs. As we will see in the next section, ground ice behavior significantly different from that of the present epoch occurs at an obliquity as low as 32° and, if we have overestimated the sublimation by as much as an order of magnitude, we need only increase the obliquity a few degrees to achieve the same ground ice behavior.

Results

In this section we present results of the behavior of ground ice under the influence of orbital evolution. We divide these results into two groups. The first consists of results from the molecular diffusion model. With it we examine the vertical distribution of ice within the regolith and how that distribution changes in time. As this model is numerically intensive, we cannot use it to...
map the expected geographic distribution of ground ice. Therefore, the second group consists of only the surface and subsurface thermal behavior. Using the molecular diffusion results as a guide, we can then map the geographic behavior of ground ice in the top few meters of the regolith.

**Molecular Diffusion Model**

Figure 3a shows a typical result for the midlatitudes (specifically, 96° × 51° north, also see Table 1; here and throughout, geographic locations are noted as longitude by north or south latitude). The figure consists of (i) the ice concentration as a function of depth within the regolith and of time for the past 1 m.y., (ii) the obliquity history for comparison, (iii) the column abundance of water in all phases within the regolith, (iv) the annual mean surface temperature compared with the mean atmospheric saturation temperature (frost point), and (v) the depth at which ice is stable with respect to the atmospheric water. Subsurface ice is stable at a depth at which the annual mean vapor-saturation density with respect to ice is equal to (or less than) the mean atmospheric vapor density (see Paper I for more discussion). This figure shows three characteristic types of behavior. First, between 1 m.y. and about 300 kyr ago, we observe a marked oscillation in the quantity of ice, and depth to the top of this ice, within the top meter or so of the regolith, ranging from nearly complete saturation to complete desiccation. This oscillation correlates precisely with the obliquity and is largely the result of oscillations in the atmospheric water abundance. As the obliquity increases, so does the atmospheric water, while the surface temperatures fall. Under these conditions, ground ice becomes stable closer to the surface and more equatorward compared to lower obliquities, and diffusion is sufficiently rapid to fill the pore space in the near-surface regolith completely with ice in just a few thousand years. Conversely, as the obliquity falls, so does the atmospheric water abundance, while the surface temperature rises. Ground ice in the near surface becomes unstable at any depth and so begins to sublime. Sublimation can be seen to remove ice down to a depth of about a meter before the obliquity completes a cycle and again begins to rise. These oscillations are also clearly observed in the regolith water column abundance (Figure 3a (ii)).

Second, during the last 300 kyr the behavior in the top meter is considerably subdued; this behavior correlates with a period of relative quiescence in the obliquity variations. Only small periodic additions of ice are seen in Figure 3a (i) and (iii) and are correlated with small changes in obliquity.

Third, during the entire history shown, the concentration of ice below a depth of about 1 m experienced a steady buildup, mostly unaffected by the large oscillations that occurred above; small oscillations can be seen in the ice density, the amplitude of which decreases with depth, while superimposed on the steady buildup. This buildup is also evident in the water column abundance shown in Figure 3a (iii), which shows a long term increase with short-term oscillations. Ice accumulates more rapidly during high obliquity than can be lost during low obliquity, the result of the direct dependence of diffusion on the vapor pressure combined with the nonlinear dependence of vapor pressure on temperature.

Comparison between the predicted depth at which ice is stable, Figure 3a (v), and the predicted occurrence of ice, Figure 3a (i), shows that during periods of high obliquity, excellent correlation exists in the top meter or so (to within the vertical resolution of the model layers). During periods of low obliquity, ice is not considered stable at any depth, yet ice persists below about a meter. By comparing the predicted depth of stability, Figure 3a (v), with the surface temperature history, Figure 3a (iv), we see that at times when ice is stable (at any depth), the mean annual surface temperature is lower than the mean frost point; the reverse is true when ice is not stable. As Figures 3a (iv) and (v) depend only on the thermal behavior, they can be useful in mapping the behavior of ice.

Another typical midlatitude result is shown in Figure 3b (120° × 33° north), which is closer to the equator than Figure 3a. Here we see the same type of behavior in the top meter of the regolith, with a similar correlation of ice abundance with obliquity and depth of stability. However, this location lacks a long term buildup of ice below a meter depth. Here, more equatorward of Figure 3a and about 10 K warmer, all the ice built up during high obliquity is lost by sublimation at low obliquity. The water column abundance in Figure 3b (iii) shows only adsorbed water with short excursions of ice condensation at high obliquity. However, despite the absence of a permanent buildup of ice, Figures 3a (iii) and 3b (iii) show approximately the same quantity of water exchanged during an obliquity cycle, about 30-40 g/cm². In addition, where in Figure 3a ice was found to be stable and present at the current epoch, in Figure 3b, ice is found to be neither currently stable nor currently present.

Figures 3a and 3b show a latitudinal dependence in the long-term behavior of ground ice between 51° and 33°. In Figures 3c and 3d we examine more extreme behavior. Figure 3c shows results for 80° south latitude. We see very little oscillatory behavior in the concentration of ground ice in the top meter of the regolith as compared to Figure 3a. This behavior is primarily due to ice being stable near to the surface continuously, with only small oscillations in the stability depth (Figure 3c (v)). At one of the warmest regions of the equator, Figure 3d (24° × 1° north) shows the other extreme in ground ice behavior. Here, ice does occur in prodigious quantities and to a depth of almost 2 m. However, not only is there no long term buildup, but the occurrence of ice is restricted to only the highest obliquities that occurred in the past 1 m.y., typically 32° or more.

Next, we examine the behavior near the geographic boundary between where we might expect to find ground ice at present and where we might not. Figures 4a through 4d show a latitudinal traverse (43°, 45°, 47°, and 49° south latitude) at 240° longitude, all with sim-
Figure 3. Typical molecular diffusion model results obtained for different latitudes. See Table 1 for locations and additional information. The results show (i) the ice concentration for depths up to 4.5 m and for the past 1 m.y. (about 532 thousand Martian years), (ii) the obliquity from Figure 1 for comparison, (iii) the total water column abundance including ice, adsorbate, and vapor, (iv) the annual mean surface temperature compared with the annual mean frost point, and (v) the depth below the surface at which ice is stable with respect to sublimation. The range of grays shown in (i) are for 0 to 100% of the pore space filled with ice. Black indicates no ice is present at that depth or time. Unless otherwise stated, models assume an initially dry regolith with 40% porosity (holding a maximum of 0.37 g/cm³ of ice), a tortuosity of 3 and a pore radius of 10 μm (see Mellon and Jakosky [1993] for details).
Table 1. Model Regolith Properties and Results

<table>
<thead>
<tr>
<th>Figure</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Thermal Inertia, (10^{-3} \text{cal/cm}^2\text{s}^{1/2}\text{K} (1/\text{m}^2\text{s}^{1/2}\text{K}))</th>
<th>Albedo</th>
<th>(T_s)</th>
<th>(T_{frost})</th>
<th>Current Ice Column, g/cm²</th>
<th>Is Ice Currently Stable?</th>
</tr>
</thead>
<tbody>
<tr>
<td>3a</td>
<td>30°</td>
<td>51° N</td>
<td>4.8 (201)</td>
<td>0.27</td>
<td>188.0</td>
<td>196.2</td>
<td>32.2</td>
<td>yes</td>
</tr>
<tr>
<td>3b</td>
<td>120°</td>
<td>33° N</td>
<td>2.5 (105)</td>
<td>0.29</td>
<td>197.6</td>
<td>198.6</td>
<td>0</td>
<td>no</td>
</tr>
<tr>
<td>3c</td>
<td>80° S</td>
<td>1° N</td>
<td>6.0 (272)</td>
<td>0.25</td>
<td>162.8</td>
<td>189.2</td>
<td>27.7</td>
<td>yes</td>
</tr>
<tr>
<td>3d</td>
<td>24°</td>
<td>51° S</td>
<td>9.1 (381)</td>
<td>0.21</td>
<td>221.4</td>
<td>198.9</td>
<td>0</td>
<td>no</td>
</tr>
<tr>
<td>4a</td>
<td>240°</td>
<td>43° S</td>
<td>5.3 (222)</td>
<td>0.31</td>
<td>195.5</td>
<td>197.7</td>
<td>0</td>
<td>no</td>
</tr>
<tr>
<td>4b</td>
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<td>45° S</td>
<td>5.1 (213)</td>
<td>0.29</td>
<td>194.5</td>
<td>197.5</td>
<td>2.8</td>
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</tr>
<tr>
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<td>47° S</td>
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<td>0.29</td>
<td>192.3</td>
<td>197.4</td>
<td>18.9</td>
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<tr>
<td>4d</td>
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<td>49° S</td>
<td>4.9 (205)</td>
<td>0.30</td>
<td>198.4</td>
<td>196.8</td>
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</tr>
<tr>
<td>5a</td>
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<td>45° S</td>
<td>6.6 (276)</td>
<td>0.28</td>
<td>196.8</td>
<td>197.4</td>
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<td>no</td>
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<tr>
<td>5b*</td>
<td>124°</td>
<td>45° S</td>
<td>6.6 (276)</td>
<td>0.28</td>
<td>196.7</td>
<td>198.0</td>
<td>7.5</td>
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<tr>
<td>5c</td>
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<td>197.6</td>
<td>198.6</td>
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<tr>
<td>5d*</td>
<td>120°</td>
<td>33° N</td>
<td>2.5 (105)</td>
<td>0.29</td>
<td>197.6</td>
<td>199.3</td>
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<tr>
<td>6a</td>
<td>96°</td>
<td>51° N</td>
<td>4.8 (201)</td>
<td>0.27</td>
<td>188.0</td>
<td>196.2</td>
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<tr>
<td>6b†</td>
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<td>51° N</td>
<td>4.8 (201)</td>
<td>0.27</td>
<td>188.0</td>
<td>196.2</td>
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<td>yes</td>
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<tr>
<td>6c†</td>
<td>96°</td>
<td>51° N</td>
<td>4.8 (201)</td>
<td>0.27</td>
<td>188.0</td>
<td>196.2</td>
<td>20.4</td>
<td>yes</td>
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<tr>
<td>6d‡</td>
<td>96°</td>
<td>51° N</td>
<td>4.8 (201)</td>
<td>0.27</td>
<td>188.0</td>
<td>196.2</td>
<td>25.2</td>
<td>yes</td>
</tr>
</tbody>
</table>

*Models were run for 2.5 m.y. of orbital history.
†The porosity was reduced from 40% to 10%.
‡The pore radius was reduced from 10 μm to 1 μm.
§The tortuosity was increased from 3 to 6.

ilar surface thermophysical properties (see Table 1). In the most equatorward of these models, Figure 4a, we see similar periodic ice formation with a depth of desiccation extending to about 1 m. A buildup of ice appears below this depth; however, the quantity is insufficient to survive long into the quiescent obliquity period during the past 300 kyr, hence no ice is currently present (nor is ice currently stable).

In Figure 4b the situation differs slightly. During the period of high-amplitude obliquity oscillations (1 m.y. to 300 kyr), Figures 4b and 4a look nearly identical, but now there is sufficient ice buildup beneath a meter to allow ice to persist to the present epoch, even though it is unstable through much of this period. The stability depth in Figure 4b (v) has changed little from Figure 4a (v); however, ice is stable at the current epoch in Figure 4b (v). We will explore this point further shortly.

Further south still, Figure 4c differs a little more. Now ice is clearly established below a meter. In addition, the desiccation observed during low obliquity does not penetrate as deeply. We also see that the total ice time is stable in the subsurface. Figure 4c (v), has increased noticeably from Figure 4a. Finally, the furthest poleward model in the sequence, Figure 4d, shows that the depth of desiccation at low obliquity has risen to less than a meter at all times. We also see in Figure 4d (v) that during most of the past 300 kyr, ice has been stable in the subsurface. A comparison of the column abundance of water between all four models shows that during the past 300 kyr the water column has decreased in Figures 4a, 4b, and 4c, while it continues to increase in Figure 4d. This behavior is caused by the substantial increase in the length of time ice is stable during this period.

How dependent these results are on the orbital history? Figure 5a shows a result similar to Figure 4a, for 124° × 40° south. Saturation/desiccation oscillations extend to greater than a meter depth with a steady buildup of ice below 2 m. As with Figure 4a, the ice built up during high amplitude obliquity oscillations sublimes away during the past 300 kyr. In Figure 5b we calculate the ground ice behavior for the same location and conditions as in Figure 5a, except we extend the model back 2.5 m.y. instead of 1 m.y. The result shows that in 2.5 m.y. there has been adequate time to build up sufficient ground ice below 2 m to survive until the present; yet, at this location, ice is unstable at the current epoch and has been unstable for the majority of the past 300 kyr. This presence of ice illustrates how ground ice can persist to the current epoch (after a 300 kyr period of quiescent obliquity) even where it is unstable. This result is not surprising, given the finite timescale for sublimation and loss of unstable ground ice. In much of the results presented, similar behavior is observed as a persistence of ground ice during low obliquities. Of equal importance, this result illustrates how the present distribution of ground ice can be influenced by its past history. In Figure 5c (which is identical to Figure 3b), we see a different behavior. This figure is an example of a model with no buildup of ice at depth, though the behavior is very similar to Figure 5a (i.e., similar ice stability and temperature histories). When we extend this model (Figure 5c) to 2.5 m.y. of orbital history (Figure 5d), we still do not see
any buildup at depth. The difference between the results at these two locations can be accounted for partly by the more equatorward model (Figures 5c and 5d) being about 1 K warmer on average and partly by the more poleward model (Figure 5a and 5h) having an annual thermal wave that propagates a factor of 2 deeper, combined with a higher amplitude temperature oscillation over orbital timescales, both thermally forcing water to a depth at which it is more difficult to be lost by sublimation. In general, Figure 5 illustrates that, while the saturation/desiccation cycles are largely unaffected by the past history of ground ice, the precise
Figure 5. The effects of the length of the orbital history near and away from the permanent ice boundary. (a and b) Are for 124° x 45° south with (a) run for 1 m.y. and (b) run for 2.5 m.y. of orbital history and illustrate a different behavior at the current epoch. (c and d) Are for 120° x 33° north with (c) run for 1 m.y. and (d) run for 2.5 m.y. (Figure 5c is the same as Figure 3b). Between Figures 5c and 5d there is no significant difference in the behavior between either history.

The present boundary of permanent ice buildup below these cycles will depend in part on the history of subsurface ice.

Next, we look at the sensitivity of our results to some of the parameters that control molecular diffusion. Figure 6a shows the model for 96° x 51° north (same as Figure 3a). In this model (as with the preceding figures) we assumed a porosity of 40%, a pore radius of 10 μm, and a tortuosity of 3 (see Paper I for a detailed discussion of these parameters). In Figure 6b we re-
Figure 6. The effects of varying diffusive parameters. All four figures show 96° x 51° north with different parameters. (a) The porosity is 40%, the tortuosity is 3, and the pore radius is 10 μm (same as Figure 3a). (b) We reduce the porosity to 10%, while (c) we reduce the pore radius to 1 μm, and (d) we increase the tortuosity to 6. The effects are discussed in the text.

duced the porosity alone from that in Figure 6a by a factor of 4, to 10%. A reduction in the porosity will both reduce the amount of ice that it takes to fill the pore space and be stored for future sublimation and reduce the diffusive flux of water vapor. The net effect is to reduce the net water column abundance and the oscillation amplitude in the column abundance by about the same factor. The water is unable to diffuse to as great a depth (below 3 m) in 1 m.y., partly because the maximum possible flux has been reduced and partly because it requires less ice to choke off the pores, restricting further diffusion; however, a longer history of diffusion will allow for condensation below 3 m. The depth of desiccation is almost entirely unchanged; the lower diffusive flux is countered by a smaller reservoir of regolith ice.

In Figure 6e we reduce the pore radius from that of Figure 6a by an order of magnitude to 1 μm. A reduction in the pore radius will reduce the Knudsen diffusion coefficient proportionally, though the range 10 μm to 1 μm spans the transition region where the pore radius is just beginning to become important (see Paper I). Here, we see that the saturation/desiccation depth is slightly reduced, a result of lower diffusive fluxes and a relatively large reservoir of pore ice. An increase in the tortuosity from 3 (Figure 6a) to 6 is illustrated in Figure 6d. The effect is the same as in Figure 6c because the reduced flux is compounded by the 40% porosity acting as a reservoir of sublimable ice during low obliquity. In short, the bulk of the quantitative and qualitative results are independent of diffusive parameters; only subtle changes are observed.

Thermal Model and Geographic Mapping

It is clear from the molecular diffusion results that significant oscillations in the behavior of ground ice
exist in the top 1 to 2 m of the regolith over orbital timescales. To better study the impact of ground ice, it would be useful to predict the geographic extent of these oscillations and the extent of permanent ice built up beneath such oscillations. Since the molecular diffusion model is too CPU intensive to allow its use in mapping the geographic behavior of ice on a point-by-point basis, we can take advantage of the correlation between the predicted stability and the predicted presence of ice within the top meter or so. By using thermal results alone, we can calculate the depth of ground ice below the surface for each $2^\circ \times 2^\circ$ bin between 60°N and 60°S for which thermal inertia and albedo have been derived [see Palluconi and Kieffer, 1981] and for each epoch for the past 1 m.y.

We begin by examining the thermal behavior of the surface and what influence the orbit has on that behavior. Figure 7 shows a map of the long-term mean surface temperature, averaged over the last 1 m.y. of orbital history. This map shows clearly that the temperatures decrease from the equator toward the midlatitudes. In the northern hemisphere we also observe cooler regions at some latitudes, particularly between about 60° and 180° and between 300° and 0° longitude. These regions are the result of large areas of combined lower thermal inertia and higher albedo as compared with adja-

Figure 7. The long-term mean surface temperature. This is an average of the annual mean surface temperatures for the past 1 m.y. (532 thousand Martian years). Maps are presented in a zero centered west longitude format and extend from 60°S to 60°N latitude.

Figure 8. The temperature deviation for (a) a high obliquity of 33.66° (730 kyr ago) and (b) a low obliquity of 14.55° (795 kyr ago). The deviations are calculated by subtracting the long-term mean temperature shown in Figure 7 from the annual mean temperatures at a given epoch. Negative values indicate cooler than average temperatures, and positive values indicate warmer than average temperatures.
cent warmer regions [see Palluconi and Kieffer, 1981]. The southern hemisphere lacks large variations in thermal inertia and albedo and, therefore, temperatures are mostly controlled by latitude.

As Figure 7 is an average over 1 m.y., the response to orbital variations has been removed. To best observe these responses, we calculate the temperature deviation, $\Delta T$, by subtracting the long-term mean temperature in Figure 7 from the annual mean temperature at an epoch. Figure 8 shows this deviation for high and low values of obliquity, respectively. For a high obliquity of 33.66°, Figure 8a shows that the surface is cooler than average (indicated by $\Delta T < 0$), with the largest cooling occurring in the north and south midlatitudes. The equatorial regions are only slightly cooler. For a low obliquity of 14.55° the opposite is observed. The midlatitudes are several degrees warmer and the equatorial regions are only slightly warmer (indicated by $\Delta T > 0$). This obliquity dependence dominates the temperature behavior between 1 m.y. and about 300 kyr ago, when the obliquity underwent relatively high amplitude oscillations.

During the past 300 kyr the obliquity has not undergone large oscillations and the $L_S$ of perihelion has dominated the temperature behavior [see also Mellon and Jakosky, 1992]. Temperature deviations in Figure 9 illustrate this behavior. In Figure 9a, perihelion occurs at an $L_S$ of 78°, near summer solstice in the northern hemisphere. The results show that surface temperatures are a few degrees cooler than average ($\Delta T < 0$) in the northern hemisphere because the short, hot northern summer allows more thermal radiation to escape to space than does the long, cool southern summer. Radiating heat to space prevents it from being seasonally stored in the subsurface and results in lower mean temperatures. In Figure 9b, perihelion occurs at an $L_S$ of 264°, near southern summer solstice (similar to the current epoch), and the resulting temperatures are warmer than average in the north ($\Delta T > 0$) and cooler than average in the south ($\Delta T < 0$). The orbital eccentricity acts to modulate this effect, maximizing the asymmetry at the highest eccentricities and reducing it to near zero for a nearly circular orbit. The structure observed in the 0 K contour is an artifact of the numerical model’s uncertainty (typically less than 0.5 K), which becomes pronounced when the temperature deviations are near 0 K.

Using the thermal results for the subsurface, we can calculate the depth at which ground ice is stable. This depth was shown in Paper I to occur where the mean annual subsurface saturation pressure with respect to ice equals the mean annual atmospheric near-surface water abundance. If this condition does not occur at any depth in the subsurface, ice is not stable. Models were run from 1 m.y. ago to the present at increments of 5 kyr, and a subset of these results will be discussed here. Figure 10 shows a time history of the depth at which ground ice is stable and its geographic extent between 655 and 620 kyr ago, when the obliquity rose from 19.55° to 33.02°. In Figure 10a the obliquity is low; the surface temperatures are warm (similar to Figure 8b), and the atmospheric water abundance is low (2.0 pr μm). As a result, ground ice stability is restricted to poleward of 60° except for a very few lo-

![Figure 9. The temperature deviation for (a) perihelion near northern summer solstice ($L_S = 78°$, 75 kyr ago) and (b) perihelion near southern summer solstice ($L_S = 264°$, 150 kyr ago). The current epoch is similar to Figure 9b. Negative values indicate cooler than average temperatures, and positive values indicate warmer than average temperatures.](image-url)
Figure 10. The depth and regional extent of ice stability for a sequence from low to high obliquity
(a) 655 kyr ago, $\Theta = 19.55$, $n_{\text{atm}} = 2.62$ pr $\mu$m; (b) 650 kyr ago, $\Theta = 21.90$, $n_{\text{atm}} = 5.01$ pr $\mu$m;
(c) 645 kyr ago, $\Theta = 24.57$, $n_{\text{atm}} = 8.8$ pr $\mu$m; (d) 640 kyr ago, $\Theta = 27.07$, $n_{\text{atm}} = 33.5$ pr $\mu$m;
(e) 635 kyr ago, $\Theta = 29.3$, $n_{\text{atm}} = 109$ pr $\mu$m; (f) 630 kyr ago, $\Theta = 31.1$, $n_{\text{atm}} = 232$ pr $\mu$m; (g) 625 kyr ago, $\Theta = 32.35$, $n_{\text{atm}} = 361$ pr $\mu$m; and (h) 620 kyr ago, $\Theta = 33.02$, $n_{\text{atm}} = 446$ pr $\mu$m.
Gray indicates ice is not stable with respect to sublimation at any depth. Contours indicate the depth to the top of the stable region; (a through d) contour intervals are 25, 50, 75, and 100 cm and (e through h) 2, 5, 10, 50, and 100 cm. The current epoch has an obliquity between Figure 10c and 10d.
cations shown; gray regions indicate no stable ice. As the obliquity increases in Figures 10b through 10h, the geographic extent of the ice stability grows equatorward as the regolith cools and the atmospheric water content rises. The depth of this stable region also moves closer to the surface; in Figure 10d the stability depth is typically much less than a meter except for very close to the boundary and two regions in the northern hemisphere between about 0° and 60° and between 220° and 300° longitude, where the thermal inertia is relatively high. The current obliquity falls between those in Figures 10c and 10d. In Figure 10e the north and south boundaries meet near 120° longitude. The large regions of shallow stability between 60° and 180° and between 300° and 0° correspond to low thermal inertia regions and lower mean temperatures (Figure 7). By 630 kyr ago in Figure 10f the only remaining regions without ice stability are the warmest in Figure 7 (approximately corresponding to the 220 K contour). At this point the atmospheric water content has reached 232 pr μm,
which saturates at about 219 K. In Figure 10g (625 kyr ago), only one 2° by 2° bin remains without stability. The obliquity has reached 32.35°. Figure 10h shows global ground ice stability, typically within 2 to 20 cm of the surface. Notice also that the stability depth is shallower in the equatorial regions than in the midlatitudes. This depth distribution is due to a combination of (1) generally lower than average thermal inertias in the equatorial regions, resulting in shallower penetration of the annual thermal wave, and (2) slightly lower than average midlatitude mean atmospheric water abundances resulting from more frequent seasonal surface saturation and water frost formation.

Figure 10 illustrates a dramatic change in the behavior of ground ice during an obliquity cycle. This behavior ranges from stable ice being restricted to the polar regions (how far poleward will be discussed below) during periods of low obliquity, to ground ice being stable globally when the obliquity exceeds about 32°. Such oscillations in the behavior of ground ice will have occurred during the high amplitude obliquity oscillations observed in Figure 1. During the past 300 kyr, however, the obliquity has been quiescent and so the atmospheric water abundance has not changed much. The \( L_S \) of perihelion has then dominated the oscillations in the geographic distribution of ice stability, by causing a geographic shift in the temperature behavior between that shown in Figures 9a and 9b. During southern summer perihelion (\( L_S \sim 270 \)) the ice stability boundaries in both hemispheres will shift north a few degrees in latitude as the south cools and the north warms (see Figure 9b). The opposite will occur during the northern summer perihelion (\( L_S \sim 90 \)).

The percentage of time over the last 1 m.y. that each location equatorward of 60° spends with ice stable at any depth is shown in Figure 11. This percentage is a combination of the length of time at high obliquity that ice is stable and the frequency at which the obliquity reaches values sufficiently high for each location. We see that equatorward of about 30° latitude, ice is stable typically 20% of the time. The percentage increases rapidly poleward of 30°, so that near 60° latitude ice is stable more than 80% of the time.

The molecular diffusion models run for various locations around the globe show that a permanent buildup of ice, below the depth of desiccation, occurs when the long-term mean frost point (long-term mean atmospheric saturation temperature) at each location exceeds the long-term mean surface temperature (see Figure 7). Figure 17 shows the expected geographic extent of permanent ice buildup based on this criterion. Black indicates no permanent ice while gray indicates the permanent buildup of ice below about 1 to 2 m. The equatorward boundary is uncertain by a few degrees of latitude due to the influences of a longer 2.5 m.y. history, as illustrated in Figure 5, and to the in-

Figure 12. The predicted distribution of permanent ice. Black indicates no permanent ice, light gray indicates both permanent ice and the stability of ice at the current epoch, and dark gray indicates permanent ice that is currently not stable. See text for more discussion.
terplay between diffusion to depths deeper than 2 m and the depth of penetration of the seasonal thermal wave. Superimposed on this map is the stability map for the current epoch (light gray). The hemispheric asymmetry in the nonoverlapping regions (dark gray) is a result of the current $L_S$ of perihelion (see above). The current stability map shown here differs in areal extent from that in Paper I due to (1) a small correction to the numerical routines and (2) the use of a spatially variable atmospheric water content. In Paper I we used an annual mean of 10 pr µm of water when accounting for surface saturation, while in the present model we use a maximum of 10 pr µm as dictated by the obliquity and equation (1) with the seasonal surface saturation described above. The combined effect is to move the current stability boundary poleward by between 0° and 4° from that of Paper I. The qualitative nature of the result is unchanged. Furthermore, interannual variability of the mean atmospheric water content, resulting in a mean that is not 10 pr µm during the current epoch, would also shift the boundaries.

So far, we have examined the thermal and ice-stability behavior between 60°N and 60°S latitude. What happens poleward of 60° and, during low obliquity, how far poleward does the region where ground ice is unstable extend? To address these questions, we calculate the stability history from pole to pole for representative high and low values of thermal inertia. Figure 13a shows the top of the stable region for the past 1 m.y. from 90°S to 90°N latitude and for a low thermal inertia of $2.5 \times 10^{-3}$ cal/cm² s¹/² K (105 J/m² s¹/² K). Depths shown extending 4 m are regions unstable with respect to ice sublimation at all depths. In the equatorial and midlatitude regions, we see the same behavior as described above; during high obliquity, ground ice is stable close to the surface and globally, while at low obliquity, ice becomes unstable. For this value of thermal inertia, ice instability does not extend further poleward than about 59°.

Figure 13b shows the same type of result for a high thermal inertia of $10 \times 10^{-3}$ cal/cm² s¹/² K (419 J/m² s¹/² K). We see the same equatorial behavior as in Fig-

![Low Thermal Inertia](image1)

![High Thermal Inertia](image2)

**Figure 13.** The top of the region where ice is stable for (a) a low thermal inertia regolith ($I = 2.5$ (105) and $A = 30\%$) and (b) a high thermal inertia regolith ($I = 10$ (419) and $A = 15\%$) from 90°S to 90°N. In the equatorial regions we see cyclic behavior, while poleward of about (a) 59° and (b) 69°, we see continuous stability.
ure 13a with the exception of the last two high obliquity periods around 400 kyr ago, when the stability does not quite reach across the equator, higher thermal inertia regions have higher mean annual temperatures and are therefore more likely to be unstable with respect to ice sublimation. In the polar regions, ice is always stable at some shallow depth, as in Figure 13a. However, for the higher thermal inertia the maximum poleward extent of the region of unstable ground ice is increased to 69°.

In Figure 13 the depth to the top of the ice in the polar regions still undergoes oscillations (as in Figure 3c), though not as large in magnitude as those regions which are more equatorward. The drop-off in depth from regions of ice stability to regions more equatorward and unstable to ice sublimation is steep, in agreement with previous studies for the current epoch [Farmer and Doms, 1979; Paige, 1989].

Discussion

In this section we will examine the implications of changes in ground-ice stability induced by orbital variations. We will focus on the nature of ground-ice stability and presence, the impact on polar cap ice deposits, and the impact on the geomorphic character of the surface.

The results of the molecular diffusion model clearly illustrate that significant quantities of ground ice can condense from atmospheric water at high obliquity, while at low obliquity ground ice will sublime, dissolating the regolith to a depth of 1 to 2 m (see Figures 3 through 6). This process is rapid enough that, within approximately the top meter, the depth at which ice is present correlates with the depth of the region within the regolith where ice is stable with respect to sublimation (i.e. transport is sufficiently rapid that ice is present where it is stable within the top meter). For this reason, depths shown in Figure 10 (about 1 m or less) should correspond to the presence of ice during an obliquity cycle; this assumes that a porous regolith exists, consistent with radar and thermal inertia observations (see Paper I and references therein).

Below the depth of desiccation (1 to 2 m) a permanent buildup of ice may exist (see Figure 12). The depth to which a permanent deposit would extend below the surface will depend on the depth of penetration of the seasonal thermal wave before it is overcome by the geothermal gradient. Water vapor will be transported deeper into the regolith where shallow temperatures are generally warmer than deeper temperatures, resulting in a vapor density gradient driving diffusive transport. While climatological oscillations in the subsurface temperature can, at times and in some locations, reverse the geothermal gradient [Mellon and Jakosky, 1992], these reversals are not capable of driving significant quantities of water vapor deep into the regolith because of the large distances and short timescales. Therefore, ice present much below the depth where the seasonal thermal oscillations die out (typically less than 10 m) must be provided by burial of existing ice or by transport of vapor upward from a ground water table [Clifford, 1991] or previously existing deeply buried ice [Fanale et al., 1986]. Only if such a non-atmospheric source exists will the permanent ice, mapped in Figure 12, be underlain by a thick deposit of ground ice.

During high obliquity, we have seen that large quantities of ice can condense within the global regolith. Since the atmosphere is assumed to be the only source of this water, it necessarily originates from the polar caps. Similarly, water lost from the regolith by sublimation during low obliquities is put back into the atmosphere and would eventually cold-trap onto the polar caps. The quantity of water exchanged in this manner is variable from location to location. In the polar regions (e.g., Figure 3c) it is small, typically less than 10 g/cm². In the midlatitudes, typically 30 to 40 g/cm² will exchange, while at the equator the exchange is as high as 50 g/cm²; the exact quantity of water exchanged with the atmosphere and polar caps at any location will clearly depend on the thermal and diffusive properties of the regolith (porosity, thermal inertia, etc.). The surface area of Mars is dominated by midlatitude and equatorial regions. If we, therefore, take 40 g/cm² as a global average, then the total amount of ground ice and adsorbate exchanged with the atmosphere and polar caps over an obliquity cycle is about 5.8 x 10¹⁵ g. If all this water were condensed onto both polar caps (poleward of 80°) during low obliquities, it would account for a deposit about 29 m thick or thicker, depending on the quantity of entrained dust or trapped gases. Interestingly, the polar layered terrain has alternating light and dark bands, thought to be dust and ice, that are typically between 14 and 46 m thick [Hlusius et al., 1987]. For comparison, Zent et al. [1986] (using a model similar to Fanale et al. [1986]) suggested that 1 to 20 g/cm² may exchange between the regolith and polar caps, possibly creating a polar deposit layer 0.7 to 14 m thick, however, their calculations did not include the equatorial regions, which dominates the magnitude and areal contribution to the total exchanged water.

At high obliquity, polar ice sublimes and condenses as ground ice in the global regolith; for a basalt, regolith ice is the dominant sink. When the obliquity decreases, the ground ice and adsorbate are released and condense back onto the polar terrain. Obviously, this process would involve much of the same water on each cycle and would produce one layer in the polar terrain. In order to create several layers a net gain would be required, such as a net transfer of water from one pole to the other pole over many obliquity cycles [e.g., Jakosky et al., 1993]. Additionally, the quantity of water exchanged in any one cycle will vary, depending on the maximum obliquity reached and the time spent at high obliquities. For example, if the maximum obliquity in a cycle was 29° (Figure 10d), then much of the equatorial region, containing a large areal fraction of the surface, would not exchange ice. The lower atmospheric water abundance at this maximum obliquity would also cause less ice to condense in the midlatitudes during the cycle. The result of this lower maximum obliquity would be less than 29 m of ice at the poles exchanged with the
regolith. The amount of water exchanged between the polar terrain and the global regolith may not be affected greatly by limits on the quantity of porous regolith material, since thermal inertia results indicate that no regions are dominated by high thermal inertia (and low porosity) bedrock on the scale mapped here [Palluconi and Kieffer, 1981].

The cyclic saturation and dessication of the top 1 to 2 m of the regolith also are likely to impact the geologic character of the surface. In our models we assume that ground ice condenses in the soil pores and that no inflation or deflation of the regolith occurs. If ground ice preferentially occupied more than the pore volume, then a cyclic inflation and deflation of the regolith would follow. This cyclic process would likely have an effect similar to that of a frost-heave cycle observed on Earth; cycles could cause small-scale features such as solifluction lobes, hummocky surfaces, and stone sorting. During times of high obliquity when ground ice is present (independent of any inflation effects), the pore ice will act to cement the regolith, which could then undergo thermal-contraction-type fracturing, possibly producing frost- or sand-wedge polygons; surface relief could form where water ice or dust and sand fills cracks opened during cooling (either seasonally or orbitally induced) and before they can close during warming. Polygonal patterns are thought to be observed at the Viking 2 Lander site [Mutlu et al., 1977], which (at about 225° x 48° north) falls just equatorward of the permanent ice boundary shown in Figure 12. Figure 11, which shows the percent of the time ice is stable (and therefore likely to be present) in the past 1 m.y., suggests that since ground ice is likely present at some times in the equatorial regions, polygons might form there, though they may be poorly developed; a latitudinal gradient in the extent and development of glacial geomorphic features is suggested by this figure. Since the depth scale of dessication of the regolith is of the order of 1 to 2 m, these geomorphic features would be expected to have a horizontal scale of the same order.

Ground ice in any quantity will choke the soil pores and restrict the diffusion of water vapor and of CO₂ desorbed from the regolith during orbitally induced thermal changes (see Paper I). While partially filled soil pores will clearly slow the exchange of CO₂ with the atmosphere, complete filling of pores can block it entirely. In regions of permanent stability, shown in Figure 12, such complete filling to a depth where the seasonal thermal oscillations die out could occur in under a hundred million years from atmospheric water alone; additionally, a deep-regolith source of water would provide for diffusion of vapor upward along the geothermal gradient, smelting ice below the seasonal oscillations and enhancing blockage. Mid-latitude and high-latitude CO₂ desorption then will be limited most likely to the top 1 to 2 m of the regolith. Care should therefore be exercised in employing adsorbed CO₂ exchange in long-term climate models (see Paper I for additional discussion).

The results presented focus on the behavior of ground ice over the past 1 to 2.5 m.y., during which time the obliquity ranged between about 15° and 35° (see Figure 1). Larger excursions in obliquity, between 0° and 60°, are also possible in the more distant past; even nonextreme values would surely enhance the effects discussed. For example, if the mean obliquity (currently near 25°) wandered to a higher value (e.g., 30°), the exchange of water with the polar layered deposits could have increased, making thicker layers, and the time which equatorial ground ice could spend modifying the surface geology would also increase. Conversely, a lower mean obliquity (e.g., 20°) might produce thinner polar layers and limit periglacial evolution. Future analysis of polar layered deposit thicknesses and nonpolar geomorphology could provide clues toward the past obliquity history during epochs of chaotic and therefore otherwise unpredictable behavior.

Future observations could provide a check for the models presented here. High-resolution imaging could allow searches for small-scale geomorphic features indicative of ground ice and its regional extent. Similar imaging of the polar layered deposits could resolve questions as to the nature of the layers and global ground-ice exchange as a mechanism of formation. Other instruments such as a gamma-ray spectrometer and neutron detector would be capable of detecting water in the top meter of the surface [e.g., Feldman et al., 1993]. More ambitious projects such as relatively shallow core drilling might locate ground ice in regions where it is currently stable or left over from an earlier epoch (see Figure 12).

Conclusions

From the results presented here we draw the following conclusions:

1. Obliquity oscillations can cause marked changes in the atmospheric water content and surface temperatures on Mars as to effect a significant global change in the stability of ground ice. Variations in the Lₜ of perihelion and eccentricity have lesser but noticeable effects. While our results encompass the past 1 to 2.5 m.y., the behavior of ground ice during these orbital oscillations should be considered representative of much of Martian history.

2. The process of molecular diffusion of atmospheric water is rapid enough on orbital timescales to produce cyclic saturation and dessication of the top 1 to 2 m of the regolith. The presence of ice in the top meter closely follows the stability of ice as predicted by these models. This correlation enables us to map the expected presence of ground ice throughout recent history. An example sequence from low to high obliquity is presented in Figure 10.

3. At obliquities of 32° or greater ground ice is stable, and likely present in the near-surface regolith, globally. This ground-ice behavior is significantly different from the current epoch, where ice stability is restricted to regions poleward of approximately 40° latitude, accounting for only about 1/3 of the Martian surface.

4. A long-term buildup of permanent ice is expected
below the depth of dessication (1 to 2 m) in some regions. This ice will extend to a depth where the seasonal thermal oscillations die out (less than 10 m); deeper ice will require nonatmospheric sources or burial of near-surface deposits. Permanent ice may persist to the current epoch in regions where ground ice is presently unstable.

5. The quantity of water exchanged between the regolith and polar caps during an obliquity cycle is variable with location and maximum obliquity. Typically, for obliquities exceeding 32°, an average of about 40 g/cm² is exchanged with the global regolith; this amount of ice is capable of creating a deposit about 29 m thick on both poles. This thickness is consistent with observations of polar-layered-deposit layer thicknesses.

6. Cyclic saturation and dessication may have a pronounced effect on the geomorphic character of the surface. Features might form such as solifuction lobes, hummocky terrain, sorted stones, and frost- and sand-wedge polygons. These features should have small horizontal scale owing to the small vertical scale of the saturation and dessication cycles. Future high-resolution images of such features might provide the best test of the models presented here.

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