MARS: THE RESPONSE OF AN ICE-RICH CRUST TO BURIAL BY A VOLATILE-POOR MANTLE. S. M. Clifford, Lunar and Planetary Institute, Houston TX 77058, USA.

Consideration of the martian valley networks, outflow channels, and various other surface features, suggests that the martian crust is water-rich and that it may host a planetary inventory of outgassed water equivalent to a global ocean 0.5 - 1 km deep [1]. However, the geologic evidence also indicates that Mars has undergone extensive resurfacing -- including the production of up to 2 km of ejecta from impacts [2], as much as 0.5 km of extrusive volcanics [3], and an unknown (but potentially large) volumetric contribution of interbedded weathering and sedimentary deposits [e.g., 4, 5]. Given the potential effect of such resurfacing on the geothermal temperature profile of the near-surface crust, and the strong influence of subsurface temperature on the stability and distribution of ground ice, the response of an initially ice-rich crust to burial by volatile-poor mantles has been considered -- including potential changes in state, distribution, and modes of volatile transport.

At the base of the cryosphere (that region of the crust where the temperature remains continuously below the freezing point of water), the added insulation provided by the emplacement of the depositional layer will result in the rise of the melting isotherm until thermal equilibrium is re-established -- resulting in the melting of any intervening ice and the subsequent drainage of meltwater into the lowermost porous regions of the crust [6, 7]. However, at the top of the ice-rich layer, the increase in local temperature will lead to the movement of the crustal 'cold-trap' into the overlying mantle -- causing the ice present at depth to be thermally redistributed until it has saturated the available pore space in the colder regolith above [7].

This redistribution may occur by three different processes that involve all three phases of water, these include: thermal vapor diffusion (where vapor migrates from the warmer depths to the colder near-surface crust), thermal liquid transport (in response to the temperature-induced gradient in soil water potential that can occur in the interfacial films between rock and ice), and regelation (the movement of ice through soil pores via pressure induced melting and refreezing) [7, 8]. Of these processes, thermal vapor diffusion is the most efficient -- its magnitude being directly proportional to both the local saturated vapor pressure (which, in turn, is a strong function of crustal temperature) and local geothermal gradient.

The potential effect of resurfacing on the thermal and volatile evolution of the near-surface crust is illustrated in Figure 1. In this particular example, the instantaneous burial of an ice-rich unit by a 100 m-thick volatile-poor mantle is considered. Because deposition is assumed to have occurred under ambient martian conditions, the initial temperature profile of the mantle is taken to be isothermal. Given a thermal conductivity of 2.0 W m^{-1} K^{-1} and a geothermal heat flux of 30 mW m^{-2}, it takes ~10^3 years for the top 300 m of the crust to thermally reequilibrate. For a mean annual surface temperature of 210 K (corresponding to a latitude of ~30°), the timescale for the redistribution of ground ice is ~10^7 - 10^8 years -- a result that assumes a mantle porosity of 20% and a characteristic pore size of 1 - 10 μm. Note that, because of the strong temperature dependence of the saturated vapor pressure of H_2O, the process of thermal vapor diffusion is significantly enhanced by the higher crustal temperatures present at depth. For example, at the base of the cryosphere, where the local temperature is near 273 K, the time required to saturate an equivalent pore volume is shorter by a factor of 10^3 (Table 1).

The results summarized in Table 1 also demonstrate the importance of the local geothermal gradient to volatile transport. Although gradients of ~0.015 K m^{-1} are thought to be representative of the crust today, models of the thermal history of Mars suggest that, 4 billion years ago, crustal gradients may have been as much as 3 - 5 times larger [9] -- implying a similar increase in the efficiency of vapor transport. On a local scale, gradients as large as 10^2 K m^{-1} may be found near active geothermal regions -- such as volcanoes, igneous intrusions, and impacts -- or on a transient basis within the diurnally- and seasonally-active layer of the martian regolith.

The results of this analysis indicate that even a small change in crustal temperature can exert a strong influence on the transport, stability, and ultimate distribution of subsurface H_2O. In particular, it demonstrates that through the process of thermal vapor diffusion, an initially volatile-poor depositional mantle, overlying an ice-rich crust, may (on a geologically short time scale) become quickly charged with ice -- a fact that may have important implications for reconciling the geologic evidence for extensive resurfacing with the geomorphic evidence for the presence of ice within the near-surface crust.

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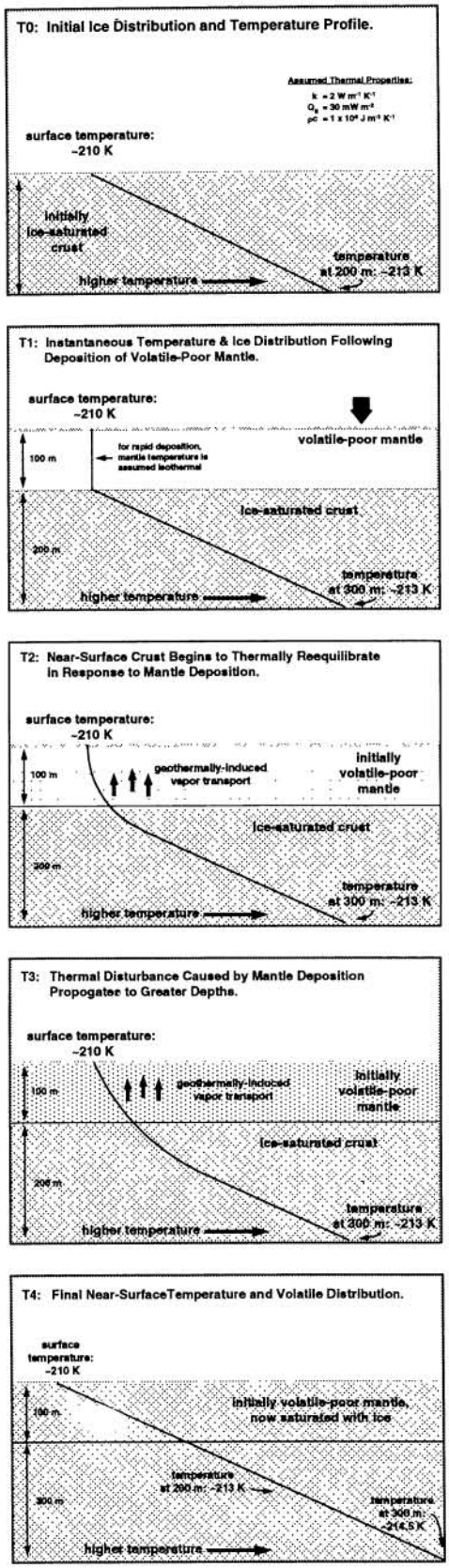

FIGURE 1 (LEFT). Time sequence illustrating the response of an ice-rich crust to burial by a volatile-poor mantle. For simplicity, both the mantle and underlying crust are assumed to have uniform values of thermal conductivity (2.0 W m^{-1} K^{-1}), porosity (20%), and pore size (1 & 10 μm). A mean annual temperature of 210 K (corresponding to ~30° latitude) and a geothermal heat flux of 30 mW m^{-2} are also assumed.

TABLE 1. TIME REQUIRED TO SATURATE A 100 M-THICK MANTLE WITH ICE (YRS)*

1 μm pore size

	Geothermal Gradient (K m^{-1})					
Temp. (K)	0.015	0.03	0.075	0.15	1	100
273	2.4E+05	1.2E+05	4.9E+04	2.4E+04	3.7E+03	3.7E+01
220	3.0E+07	1.5E+07	5.9E+06	3.0E+06	4.4E+05	4.4E+03
210	1.0E+08	5.0E+07	2.0E+07	1.0E+07	1.5E+06	1.5E+04
200	3.9E+08	2.0E+08	7.8E+07	3.9E+07	5.9E+06	5.9E+04
190	1.8E+09	8.8E+08	3.5E+08	1.8E+08	2.6E+07	2.6E+05
180	9.5E+09	4.7E+09	1.9E+09	9.5E+08	1.4E+08	1.4E+06
170	6.3E+10	3.1E+10	1.3E+10	6.3E+09	9.4E+08	9.4E+06
160	5.3E+11	2.6E+11	1.1E+11	5.3E+10	7.9E+09	7.9E+07

10 μm pore size

	Geothermal Gradient (K m^{-1})					
Temp. (K)	0.015	0.03	0.075	0.15	1	100
273	7.1E+04	3.6E+04	1.4E+04	7.1E+03	1.1E+03	1.1E+01
220	6.9E+06	3.4E+06	1.4E+06	6.9E+05	1.0E+05	1.0E+03
210	2.4E+07	1.2E+07	4.8E+06	2.4E+06	3.6E+05	3.6E+03
200	9.5E+07	4.7E+07	1.9E+07	9.5E+06	1.4E+06	1.4E+04
190	4.4E+08	2.2E+08	8.8E+07	4.4E+07	6.6E+06	6.6E+04
180	2.4E+09	1.2E+09	4.8E+08	2.4E+08	3.6E+07	3.6E+05
170	1.6E+10	8.2E+09	3.3E+09	1.6E+09	2.5E+08	2.5E+06
160	1.4E+11	7.2E+10	2.9E+10	1.4E+10	2.2E+09	2.2E+07

*Assumed porosity of 20%

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