Chapter 1

Subsurface Ice on Mars

Water has played a central role in the evolution of surface materials (*e.g., Banin et al.*, 1992), geomorphology (*e.g., Baker et al.*, 1992; *Baker*, 2001), and in potentially habitable environments on Mars (*e.g., McKay et al.*, 1992; *Cabrol et al.*, 2007). Areas of ancient Mars experienced massive floods capable of sculpting large channels, fluvial valleys, and dendritic networks (*e.g., Sharp and Malin*, 1975; *Carr*, 1981; *Baker*, 1982; *Baker et al.*, 1992; *Carr*, 1996; *Masson et al.*, 2001; *Komatsu and Baker*, 2007). Formerly wet environments have produced characteristic local and regional morphologies (*e.g., Banin et al.*, 1992), and various mineral assemblages including iron oxides (*e.g., Christensen et al.*, 2000), sulfate deposits (*e.g., Squyres and Knoll*, 2005), and diagenetic alteration products (*e.g., Squyres et al.*, 2004b). Though not as active in forming surface features today as in the past, water continues to modify the planetary surface through its role in the formation of the polar caps, gullies, and possibly slope-streaks (*e.g., Jakosky and Haberle*, 1992; *Carr*, 1996; *Malin and Edgett*, 2000; *Mellon and Phillips*, 2001; *Ferris et al.*, 2002; *Schorghofer et al.*, 2002).

The water cycle on Mars includes a number of reservoirs which may exchange with each other on various timescales. The atmosphere, polar caps, and near-surface regolith may interact through convective transport, condensation/sublimation, and adsorption (*e.g., Farmer and Doms*, 1979; *Jakosky*, 1983; *Clifford*, 1993; *Smith*, 2002). Perhaps the most voluminous reservoir, however, is the porous regolith (*Rossbacher and Judson*, 1981), which presently communicates with the atmosphere through the physical process of diffusion. The timescales involved in this exchange range from diurnal breathing to multi-million year climate variation. The rate of water vapor diffusion through the regolith determines how long ice can survive in the subsurface when exposed to a drier atmosphere (*e.g., Smoluchowski*, 1968; *Farmer and Doms*, 1979; *Paige*, 1992; *Mellon and Jakosky*, 1993; *Mellon et al.*, 2004; *Schorghofer and Aharonson*, 2005). Diffusion also determines how fast the regolith can be recharged with atmospherically derived vapor (*Mellon and Jakosky*, 1995; *Hudson et al.*, 2008) subject to the conditions of vapor supply and vapor gradients. Predicting present and past reservoir locations and volumes, constraining exchange rates between them, and understanding the dynamics of subsurface ice evolution motivate the study of the diffusion of water vapor in the subsurface of

Mars.

Extensive subsurface ice on present-day Mars has been observed within a meter of the surface at latitudes polewards of approximately 60° (*Boynton et al.*, 2002; *Feldman et al.*, 2002). Past epochs on Mars no doubt also harbored buried ice, though its distribution may have been significantly different (*Squyres et al.*, 1992; *Mellon and Jakosky*, 1995; *Carr*, 1996). A number of investigations have examined the equilibrium behavior of subsurface ice under the present climate (*e.g., Mellon and Jakosky*, 1993; *Mellon et al.*, 2004; *Schorghofer and Aharonson*, 2005), and under various climate conditions as modulated by orbital and axial parameters (*e.g., Mellon and Jakosky*, 1995; *Mellon et al.*, 1997; *Levrard et al.*, 2005; *Schorghofer*, 2007). The ice, though protected from extremes of temperature by the overlying regolith, diffusively exchanges with the atmosphere and feels the effects of atmospheric vapor content and subsurface temperature. While unrelated to the equilibrium position of the ice table, the diffusive properties of the porous media overlying ice exert a first-order control on the rate of response of the ice table's position to changing conditions.

Sub-freezing (*i.e.,* bulk liquid-free) vapor diffusion in soils finds further application in the evolution of subsurface ice on Earth. The longevity of ice in cold, arid locations such as the Dry Valleys of Antarctica constrains timescales of terrestrial climate change (*Sugden et al.*, 1995; *Hindmarsh et al.*, 1998). Many sites in the McMurdo Dry Valleys have dry permafrost within the upper meter of the glacial till comprising the valley floor (*Bockheim*, 2002). The depth and extent of this ice may be controlled in part by vapor diffusion processes, particularly where precipitation and mean annual temperatures are low and liquid water is largely absent (*Hindmarsh et al.*, 1998; *McKay et al.*, 1998; *Schorghofer*, 2005). While the studies in this work are motivated by Mars, they are also be relevant for such localities on Earth.

1.1 The Ice Table

When diffusive processes govern the distribution of ground ice, the interacting driving forces of temperature and vapor density control its distribution. The vapor concentration gradients exist between the ice and the atmosphere, and also within the ice itself if porous. In a steady-state environment, the interface between a perennially ice-free regolith and a perennially ice-bearing regolith defines the physical ice table. The temporal qualifier allows the regolith above the ice table to experience periodic ice deposition and loss in response to daily and seasonal changes of atmospheric humidity and subsurface temperature. Beneath such regions that harbor no annually stable ice, the balance of driving forces creates a finite depth of equilibrium: the depth where the mean annual vapor density of the atmosphere matches the equilibrium vapor density over ice. When the physical ice table is not coincident with this equilibrium depth it evolve toward it by gaining or losing mass.

1.1.1 Observational data

Recent measurements by the Gamma Ray Spectrometer on the Mars Odyssey spacecraft have been used to identify significant quantities of subsurface hydrogen in the upper meter of the martian regolith (*Boynton et al.*, 2002; *Feldman et al.*, 2002). In equatorial regions, the hydrogen-rich areas are usually less than 10% H by mass, which may be due to hydrated minerals or adsorbed water (*Feldman et al.*, 2002; *Feldman et al.*, 2004a; *Jakosky et al.*, 2005). However, mid- to high-latitude regions exhibit quantities of hydrogen which cannot be accomodated by any known mineral but one: water ice. Observed hydrogen abundances converted to water-equivalent units reveal quantities up to 50% water by mass (70% water ice by volume) in certain areas. The black contours in Figure 1.1 show the polar distribution of these measured water contents in units of mass percent. Some of this ice may fill the pore spaces of a particulate medium, while some may exist as layers of relatively debris-free ice; the existing data are insufficient to determine an unambiguous depth distribution. Different regions may prove to harbor different patterns of ice density with depth and may also be heterogeneous in ice content over small spatial scales.

1.1.2 Equilibrium models

Surface water was abundant in Mars' ancient past and produced numerous large features on the surface such as chaotic terrain, outflow channels, and drainage networks (*Carr*, 1996). In more recent geological times, most of the remaining water has retreated into the subsurface as ice (*Squyres*, 1984), and some has formed the permanent ice caps and polar layered terrain. Early models of the evolution of subsurface ice investigated the retreat rates and present depths of buried ice sheets which had been sublimating into a dry atmosphere for billions of years (*Smoluchowski*, 1968; *Soderblom and Wenner*, 1978). Theoretical understanding of astronomically induced changes to Mars' climate have brought forth the modern paradigm that significant fluctuations of insolation distribution and atmospheric water content have taken place on timescales of millions of years, or even faster (*Ward*, 1974; *Toon et al.*, 1980; *Carr*, 1996). This has led to a suite of models which predict the current equilibrium position of the ice table given knowledge of the present day atmospheric water content, insolation, and surface thermal properties. The degree to which such models accord with presently available or future observations may be explained in part by the rate at which exchange between this reservoir and the atmosphere can occur.

Mellon and Jakosky (1993) were the first to employ a complete vapor diffusion model to determine the global equilibrium positions of the ice table. Improved measurements of soil thermal properties and the effect of pore-ice on regolith thermal conductivity were incorporated into a later implementation of this model (*Mellon et al.*, 2004). *Schorghofer and Aharonson* (2005) have employed a similar model to create maps of subsurface ice stability which include regions of seasonally stable and permanent ice. Figure 1.1 shows the results of their model, compared to the data from the Mars Odyssey spectrometers.

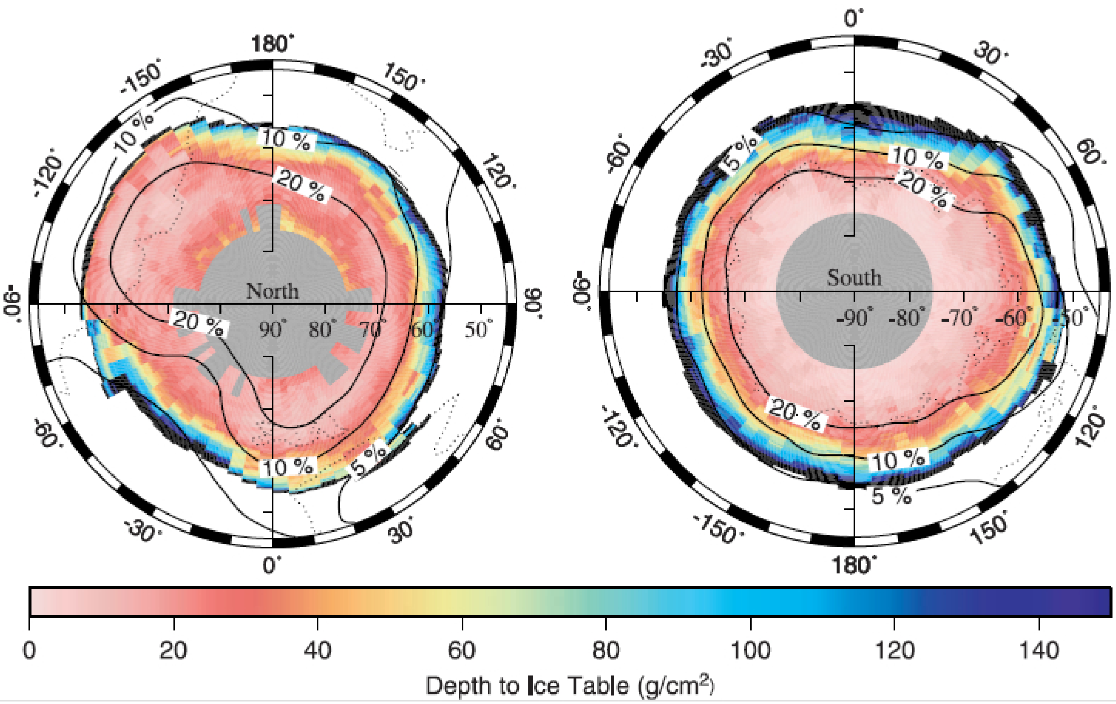

Figure 1.1: Polar maps of predicted ice table burial depths in gm cm^{-2} (color scale), and mass percent of water-equivalent hydrogen derived from Mars Odyssey measurements (solid black contours) *Feldman et al.* (2003). Ground ice is unstable in the white areas, and missing data are shown in gray. The assumed volumetric ice fraction is 40%. Dashed lines are 200 J m^{-2} K^{-1} s$^{-1/2}$ contours of thermal inertia. Figure reproduced from *Schorghofer and Aharonson* (2005).

Models of modern subsurface ice equilibrium and observations of real subsurface ice match quite well, strongly suggesting that much subsurface ice resides at or close to its equilibrium position. This further implies either a climate which has been relatively stable for some time, a fast response time of subsurface ice distribution, or perhaps both.

Several effects may be responsible for deviations of the observations from model predictions of ice position and quantity. Shallow regolith thermal properties may be probed with diurnal temperature measurements to derive near-surface thermal inertias; however, the thermal structure for much of Mars beneath a few centimeters remains hidden. Just as the thermal properties of the near-surface regolith varies significantly across the surface of Mars (*Mellon et al.*, 2000), so may the deeper subsurface. An explanation for the mismatch central to the theme of this work is the moderation by overlying ice-free regolith of the response time of the ice to a changing equilibrium. As with its thermal properties, the martian regolith is no doubt heterogeneous in its geometric structure as well; barriers in different regions will possess differing resistances. Ice in one location may be able to respond quickly to a large shift in equilibrium position, while a more shielded spot would respond sluggishly. Such heterogeneities may exist over local scales, while large regions of similar geologic character may exhibit broadly similar rates of response.

1.2 Regolith Breathing

The position of the equilibrium ice table depends only on the temperature of the subsurface ice and the vapor content of the atmosphere. As diurnal, annual, and obliquity-scale temperature waves propagate into the subsurface, and as the atmospheric moisture content changes, they alter the regions of ice stability. Rather than affecting this equilibrium, the physical process of diffusion governs the *rate* at which vapor exchange occurs. Thus the 'breathing' of the regolith, particularly on timescales longer than annual, is subject to first-order control by the process of molecular diffusion, parameterized by the diffusion coefficient.

1.2.1 The diffusion coefficient

The magnitude of mass flux driven by a concentration gradient is proportional to that gradient, a relationship called Fick's First Law of Diffusion. The diffusion coefficient, with units of length squared per time, is the proportionality constant relating the gradient and the flux. The value of this quantity for two different gases in unconstrained space depends on interactions between molecules and is understood in terms of kinetic theory (*Chapman and Cowling*, 1970). When a porous medium occupies some of the volume through which diffusion occurs, interactions between molecules and the material reduce this free-gas diffusivity by an amount which depends on the geometry of the space available to gas transport.

Figure 1.2 schematically shows the vapor exchange interaction between a ground ice and the atmosphere, moderated by an intervening layer of porous ice-free regolith. The schematic ignores the effects of adsorbed water, which are unimportant on timescales of one year or longer (see Section 3.6.1 and *Schorghofer and Aharonson*, 2005). The net flux of vapor moving either in or out of the regolith (but never both at once) is indicated by the wavy arrows. The gradient of water vapor concentration in the host gas (CO_2 on Mars) drives this transport. This gradient exists between the the atmosphere and the shallowest ice. But diffusion-driving gradients can also exist amongst various depths within the ground ice if it is porous. Beneath the deepest ground ice, vapor can diffuse into the regolith against the geothermal gradient if it remains below the saturation density. If ice deposits in this region, the vapor pressure of deeper, warmer ice will be greater than the shallower ice above, and the mass flux through ice-bearing layers will always be toward the atmosphere.

Large diffusion coefficients reduce mass flux relative to that of free-gas diffusion only slightly; a low diffusivity reduces the flux and will impede all vapor transport if the diffusion coefficient is identically zero. Values of the diffusion coefficient for the martian regolith have been used by a

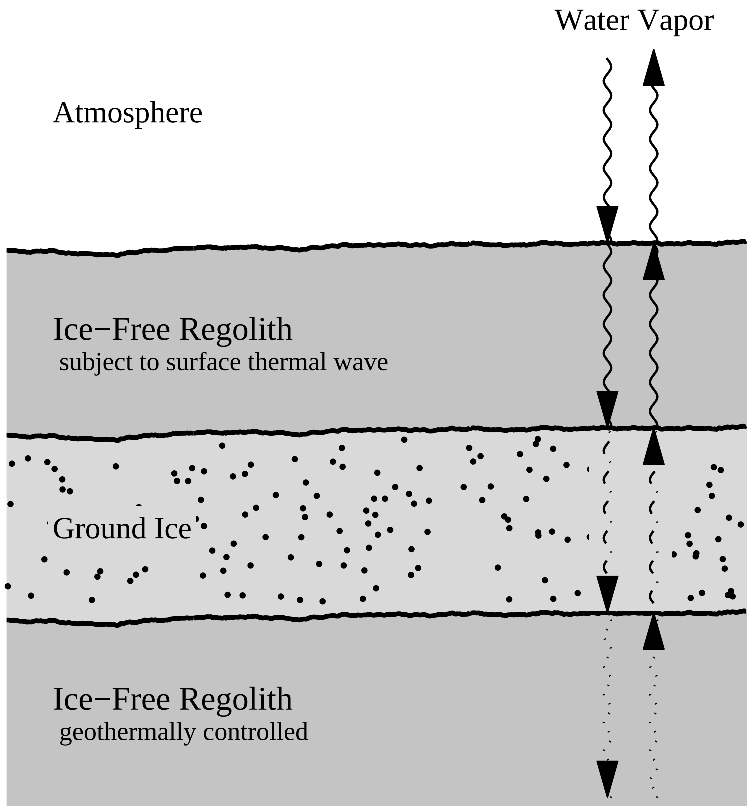

Figure 1.2: Schematic of various diffusion regions in an ice-bearing regolith. The mean annual vapor density gradient may be directed into or out of the upper layers of the regolith and the next flux may be directed inward or outward, though never both at once. Ground ice may fall on a continuum which includes pure slab ice, pore-filling ice, and ice partially filling pore spaces. Only if the ice-bearing medium is porous will diffusion occur in this layer. In the bottom layer, where temperatures are dominated by the geothermal gradient, the flux of water may still occur in either direction if unsaturated, but will always be directed toward the atmosphere in a saturated (*i.e.,* ice-bearing) region. Note that both the upper and lower boundaries of the ground ice are not static.

number of previous investigators as a parameter in models of subsurface vapor communication with the atmosphere. These values are based both on predictions of theoretical work and extrapolations of experimental diffusion data to Mars surface conditions. No samples of the martian regolith, much less undisturbed samples with their natural pore structures intact, are available and *in-situ* measurements on Mars thus far have not addressed soil diffusivities directly.

1.2.2 Previous investigations

Many of the papers described in this section calculate diffusion coefficients from the kinetic theory of gases as applied to porous media. All such methods require additional input parameters describing the geometry of the pore space. Porosity, or void volume, may be estimated from particle size distributions and shapes or measured directly for real samples. Another parameter, tortuosity, which describes the degree of interconnectivity and convolution of the void space, is not amenable to direct measurement. Even for mathematically simulated pore spaces, the value of this dimensionless number depends on the physical model used and proves intractable to calculate in all but the most simple cases. No robust relationship between porosity and tortuosity exists that holds for a wide variety of particle shapes and sizes. Thus, the value of tortuosity is usually obtained from experiment by measuring porosity and diffusivity. The ratio of porosity to tortuosity parameterizes the change in diffusion coefficient from free-gas to porous-media conditions and is often called the obstruction factor.

In a seminal paper, *Smoluchowski* (1968) calculates diffusion coefficients for a range of particle sizes and porosities using the kinetic theory of gases as presented by *Kennard* (1938) and *Evans et al.* (1961). He presents a range of values between 4×10^{-4} and 12 cm^2 s^{-1}. The lower end of this range applies to porosities of 1–10% and particle diameters less than 10 μm. Such values may be appropriate for deep, compacted regoliths or for porous rock such as sandstone. For more typical near-surface, unconsolidated soil properties, *i.e.,* porosities of 50–80% and 10–200 μm particle diameters, he obtains diffusivities between 0.7 and 11 cm^2 s^{-1}. Smoluchowski cites no reference for tortuosity, but presents values of 1, 5, and 10, for 80%, 50%, and lower porosities, respectively.

Rather than strictly calculating from theory, *Flasar and Goody* (1976) directly reference experimental measurements of gas diffusion in porous media. The works cited therein include *Currie* (1960), who performed experiments on diffusion through dry granular materials of hydrogen in air at STP, and also *Papendick and Runkles* (1965), whose work considered oxygen diffusion in porous media at STP. Both of these earlier works present factors relating free-gas and porous media diffusion coefficients for a variety of materials. Flasar and Goody use these to determine a range of porous-media diffusion coefficients of 0.4–13.6 cm^2 s^{-1} for the surface of Mars at 610 Pa and 210 K. The upper limit they give is the temperature- and pressure-extrapolated value of free-gas diffusion given by Boyton and Brattain in *Washburn et al.* (2003) (but see Section 2.3). A real porous medium will have an obstruction factor less than unity, hence the diffusion coefficient would necessarily all below this limit.

Farmer (1976) cites both Smoluchowski and Evans *et al.,* and quotes Smoluchowski's values directly. Farmer shows that his model, with a regolith diffusivity of 1.0 cm^2 s^{-1}, matches the observed amplitude of martian seasonal variation in atmospheric water content at 25°N latitude, and presents a range of 1.0–3.0 cm^2 s^{-1} as appropriate for Mars. *Jakosky* (1983) references Flasar and Goody but gives a slightly modified range of 0.3–10 cm^2 s^{-1}.

Clifford and Hillel (1983) examine the mechanics of diffusion on Mars in great detail; they also cite measurements of diffusion in porous media, all performed at 25–30°C. Most involve the diffusion of H_2 or He, and many were performed at pressures of 1 atmosphere or higher. Their derivation from kinetic theory computes values for diffusivity given only a pore size distribution and a value for tortuosity.

Fanale et al. (1986) and *Zent et al.* (1986), in tandem papers, perform their own calculations to derive diffusivity from gas kinetics and reference Smoluchowski for values of tortuosity. They use a porosity of 50% for all cases and a tortuosity of 5. For a pressure of 610 Pa and a temperature of 210 K, their expression gives a free-gas diffusivity of 13.2 cm^2 s^{-1} and a porous medium diffusivity of 0.44 cm^2 s^{-1}. These values are essentially the same as the limits given by *Flasar and Goody* (1976).

A later paper by *Clifford and Hillel* (1986) focuses on Knudsen diffusion in the Mars regolith (*i.e.,* diffusion through small pore spaces in the limit of molecule–wall collisions). *Fanale et al.* (1986) estimate a diffusion coefficient of 0.02–0.22 cm^2 s^{-1} for pore radii from 1 to 10 μm.

Mellon and Jakosky (1993) present a diffusive model for the regolith. This work includes a detailed derivation of the porous media diffusion coefficient similar to *Clifford and Hillel* (1983), but uses collision integrals to calculate free-gas diffusion coefficients, while Clifford and Hillel cite *Wallace and Sagan* (1979). For a pressure of 600 Pa, a temperature of 200 K, and a pore radius of 1–10 μm, the expression of Mellon and Jakosky gives a porous medium diffusion coefficient of 2–10 cm^2 s^{-1}.

The experiments performed or referenced in the above mentioned investigations were conducted above the freezing point of water and at high (relative to present-day Mars) pressures. Few measurements have focused on the diffusion of water vapor, a substance condensible in the temperature range of interest and highly adsorptive on most natural materials. The measurements of free-gas diffusion of H_2O in CO_2 at conditions other than those appropriate to Mars' surface have been extrapolated in temperature and pressure. Only recently have direct measurements of the diffusion of gases in porous media at conditions of low temperature and Mars surface pressures begun to appear in the literature (*Chittenden et al.*, 2006; *Bryson et al.*, 2007). The experiments described in Chapters 3 and 4 measure the diffusion coefficient of a variety of porous materials. The physical configuration of these experiments are chosen to mimic the temperatures, pressures, humidities, and diffusing gas species occurring on the surface of present-day Mars.

1.3 Barriers to Diffusion

Most experiments providing data on granular media diffusivities have been conducted with particles of uniform size, a circumstance which limits their applicability to natural particle assemblages. As demonstrated by all landed Mars missions, the surface of Mars exhibits a wide variety of soil types with different particle sizes, morphologies, and compositions (*Yen et al.*, 2005), and small regions exhibit heterogeneous distribution of these types. This suggests that the diffusive properties of the martian regolith may vary over small length scales.

On Earth, the mobilization of chemical species by liquid water may produce horizons of mineralization which have significantly reduced porosities and permeabilities relative to the host rock. Such barriers impede the movement of pore water and can form perched water tables and aquacludes. Though perhaps not mobile in the present climate, high concentrations of subsurface salts detected by the MER spacecraft (*Vaniman et al.*, 2004; *Yen et al.*, 2005), and the common occurrence of surface duricrusts (*Landis et al.*, 2004; *Wang et al.*, 2006), admits the possibility that liquid water has redistributed and redeposited such soluble minerals within regolith pore spaces. If the crystallization of these salts tended toward a pore-filling habit, rather than forcing the grains apart, the reduction of pore space would be coincident with a reduction in diffusivity and reduced water vapor flux. Barriers may persist indefinitely in subsequent drier conditions and impede the flow of water vapor, even if the salt minerals responsible for pore closure are water soluble.

Considering physical rather than chemical processes, the regolith of Mars exhibits a wide variety of particle sizes, from milimeter-sized and larger grains to aeolian sediment to dust fine enough for atmospherically suspended (*Kahn et al.*, 1992; *Squyres et al.*, 2004a). Aeolian processes can sort grain sizes such that a near-surface regolith could exhibit an particle assemblage of uniform size and shape (*Sullivan et al.*, 2005). Yet many processes also exist which can mix sediments, giving rise to soils with a broad particle size distribution. Fine particles deposited from atmospheric suspension may fill in the pores of a coarser matrix, reducing the diffusivity of the bulk material (*Farmer*, 1976). Alternatively, dust may also coat larger grains and prevent close packing, 'fluffing' the regolith, increasing the overall porosity and thereby increasing the diffusion coefficient.

Dust alone may comprise some near-surface regolith in various regions of Mars. Such dust, with particle sizes on the order of a few microns or less, can exhibit either a high or low diffusivity. If physically compacted or dispersed in a lubricating medium such as water, fine dust can achieve close-packed arrangements which may have pore sizes comparable to the particle diameters. A broad-scale overburden will encourage mechanical compaction of dust also producing small pores and thus diffusivities. Processes involved in the generation of polar layered terrain may produce low diffusivity dust layers in the following manner: Dust-bearing ice is deposited during one season or climate regime and then lost during the other half of the cycle. During the phase of net loss, as the ice sublimates, the remaining dust forms a lag. While this lag may initially have a high diffusivity, subsequent deposition of ice above would compact in, increasing its ability to preserve the ice beneath during future periods of loss (*Levrard et al.*, 2005).

A number of factors act to prevent dust derived from airfall (in the absence of snow) from packing closely. The low mass of dust particles (implying low kinetic energies) and the inefficiency of chemical weathering processes on Mars permit jagged edges to persist. The irregular dust grains will interlock more readily than smooth particles, and resist falling into compact arrangements. Electrostatic charging of dust should be efficient on Mars (*Merrison et al.*, 2004), and airborne particles may flocculate into aggregates which retain their low density upon deposition (*Abrahamson and Marshall*, 2002). These effects may act alone or in concert to produce a larger diffusivity than highly compacted dust with micron-scale pores.

This work includes diffusion coefficient measurements for media of known dust fraction, salt content, or state of compaction. The results assist understanding of how the diffusivity of a given natural porous medium on Mars may be modified by these factors.

1.4 Growth of Subsurface Ice

In the present climate, ice exists in the shallow subsurface at latitudes poleward of about 60 degrees as predicted by models (*e.g., Leighton and Murray*, 1966; *Mellon and Jakosky*, 1993; *Schorghofer and Aharonson*, 2005) and confirmed by observations (*e.g., Boynton et al.*, 2002). *Litvak et al.* (2006), show that the ice content of the ice-rich layer depends on latitude, and the ice content they derive is consistent with the existence of interstitial ice in some latitude range. During previous epochs, the stability regions were hypothesized to occupy higher latitudes when Mars' obliquity was low, and may have been global during high obliquity periods (*Mellon and Jakosky*, 1995; *Head et al.*, 2003). While some of the presently observed ice may have been emplaced by precipitation and subsequent burial, much of the shallow ground ice has been derived directly from atmospheric water vapor (*Mellon and Jakosky*, 1993; *Schorghofer*, 2007).

Ice accumulates in a permeable medium if a humidity gradient supplies water molecules to locations where the gas-phase vapor density exceeds saturation. On Mars, such a gradient can exist between a warm and humid but unsaturated atmosphere and a cold and therefore saturated location in the subsurface; the flux being directed toward the already saturated cold spot. This infilling phenomenon was first modeled numerically by *Mellon and Jakosky* (1993) and these models indicate that, for latitudes where stability criteria permit subsurface ice, it will accumulate in pore spaces beneath an ice-free layer.

Theoretical treatment of the equilibrium ice table suggests that the interface will be sharp, transitioning from an ice-rich regolith to an ice-free one over short distances (*Schorghofer and Aharonson*, 2005). The boundary for a given atmospheric water content will follow subsurface isotherms and will therefore be subject to perturbations from inhomogeneities in surface albedo and the presence of masses of relatively higher or lower thermal inertia (*Sizemore and Mellon*, 2006).

The growth of ice in a porous medium been treated theoretically (*Hobbs and Mason*, 1964; *Hobbs*, 1974). From surface energy arguments, ice is expected to form first at grain contact points, *i.e.,* areas of high negative curvature. However, the phenomenon of diffusive ice growth in a porous medium has not been previously demonstrated experimentally. The distribution of accumulating ice feeds back into the geometry of the pore spaces and also the thermal conductivity of the bulk medium.

The formation of subsurface ice reduces a porous medium's diffusivity via constriction of the pore space. As ice accumulates, it reduces the available volume for vapor transport and thereby diminishes the mass flux. A regolith with an initially large pore diameter will transition from Fickian-dominated to Knudsen-dominated diffusion as the open pores become smaller. Subsurface ice grows most rapidly near the equilibrium depth (*Mellon and Jakosky*, 1995; *Schorghofer and Aharonson*, 2005), so constriction will be most efficient at this point, strongly inhibiting diffusion and reducing deposition rates when filling fractions are high.

Laboratory experiments described herein produce for the first time the phenomenon of diffusive transport and ice deposition in a porous medium at Mars conditions. The resulting ice content profiles reveal the presence and sharpness of the developed ice table and its depth relative to the predicted equilibrium position. A variety of atmospheric humidity regimes and experiment durations are employed to provide information about the rate at which the pores fill; long-duration experiments allow the ice-filled state to be approached. A numerical model of ice deposition, run with experimental temperatures, humidities, and ice-free diffusivity as inputs, incorporates parameterizations of pore-space reduction. Comparisons between the model and experimental data allow these schema to be evaluated.

1.5 Subsurface Ice Evolution

Much of the predictive work concerning subsurface Mars ice has focused on the ice table and its equilibrium position given the current Mars climate (*Leighton and Murray*, 1966; *Farmer and Doms*, 1979; *Fanale et al.*, 1986; *Paige*, 1992; *Mellon and Jakosky*, 1993, 1995; *Mellon et al.*, 2004; *Schorghofer and Aharonson*, 2005). There are only a few investigations of the behavior of subsurface ground ice on Mars beneath the ice table. Bounded above by this interface, the region termed the cryosphere experiences perennial ground ice, analogous to permafrost on Earth. The quantity, depth, and distribution of ice beneath the ice table affects estimations of total Mars water inventories, the the interpretation of observations by planned and future missions, and our understanding of the present and past state of water on Mars.

Clifford (1993) extensively discusses the subsurface hydrology of Mars under the assumption that a cryosphere of varying thickness overlies an active hydrosphere which can supply water from deep underground to shallower depths. This recharge allows ground ice to exist in the warm and otherwise dry equatorial regions without invoking climate change or extreme preservation mechanisms. *Mellon et al.* (1997) consider the evolution of a subsurface ice sheet which initially saturates 200 m of regolith pore space. Because of the depth range covered, only the geothermal gradient drives diffusive flux through this model and, as in *Clifford* (1993), the effect of the annual thermal wave on ice evolution in the shallow regolith is not considered. Such shallow depths, however, are a region of the cryosphere easily observable by landed missions and therefore invites particular consideration.

The Mars Odyssey detection of high-latitude subsurface hydrogen, interpreted as ice, is limited to the upper meter of the regolith column. Direct investigations by the Phoenix Mars Lander, and perhaps future landed missions as well, will probe the shallow non-polar subsurface. Models which track the evolution of ice and predict its subsurface distribution in this depth range will help direct future measurements and will aid the interpretation of observations.

In *Mellon and Jakosky* (1995), the authors explore the distribution of ice derived from atmospheric water vapor in the upper few meters of the regolith at various latitudes for 2.5 Myr of Mars' orbital history. For latitudes equatorward of 60–70 degrees, their model shows that shallow regolith desiccation could occur to a depth of 1 to 2 meters. This is consistent with the work of *Schorghofer* (2007), who models the evolution of a massive ice sheet for 5 million years, incorporating the formation of a diffusive lag and ice redeposition in pore spaces. The simulation by *Schorghofer* that includes variable atmospheric humidity reveals three significant loss events between 300–600 thousand years ago. This implies that the subsurface ice targeted by Phoenix and much of the high-latitude ice sensed by Mars Odyssey has formed as a pore-filling substance since that time.

The amount of ice able to accumulate since the last major loss event will depend on feedbacks of deposited ice on the regolith diffusion coefficient and subsurface thermal conductivity, in addition to the availability of water. The shallowest ice-bearing regions (*i.e.,* the ice table) accumulate ice most rapidly and therefore exhibit greater reduction of flux than deeper, less ice-rich levels. Investigations by *Hudson et al.* (2008) (Chapter 5) indicate that constriction deviates from a linear function in filling fraction, enhancing this effect and resulting in a reduced quantity of deposited ice for similar temperature histories.

Ice within a particulate regolith also modifies the thermal conductivity of the bulk material (*Paige*, 1992), altering the amplitude and penetration depth of the thermal wave. Both the ice which deposits in pore spaces from atmospheric vapor and deep relic ice left behind after the last climatically induced loss event will contribute greater conduction of heat to depth. The effect of large-amplitude temperature variations on ice temperature will be strongest for ice within a few diurnal skin depths of the surface. Many regions with subsurface ice today exhibit a layer of ice-free regolith which protects shallow subsurface ice from these extremes, yet at some locations (*e.g.,* the Phoenix landing site) the thinness of the barrier may allow attenuated diurnal fluctuations to penetrate to the stable ice.

The feedback of these mechanisms and Mars' complex insolation history preclude analytical solutions to ice distributions over long timescales. Forward numerical modeling under assumptions of identical insolation and atmospheric humidity history will permit the effects discussed above to be examined independently. The present-day ice content predicted by the models will differ depending on the position of deep ice, the thickness of shallow ice-free layers, and the degree of non-linearity involved in constriction. The magnitude and direction of the effects can thus be assessed and this understanding will assist interpretations of shallow ice content profiles observed on Mars in the future.

1.6 Implications for Future Investigations

Future studies will employ experimental data such as those reported here to produce more accurate models, and will use the results of these laboratory investigations and simulations of ice growth and loss processes to inform interpretations of observational data. These possibilities have motivated this study of porous media diffusivities and of the evolution of subsurface ice at Mars surface conditions.

1.6.1 Global climate models

The water cycle of Mars has been an object of investigation using sophisticated general circulation models (GCMs) incorporating a variety of physical processes (*Forget et al.*, 1999; *Richardson and Wilson*, 2002; *Haberle et al.*, 2003). The latest generation of these models now include vapor diffusion processes and permit the subsurface to be both a source and sink of water vapor. While the amount of water potentially evolved from any given reservoir in one year's time under the present climate is ∼10–100 precipitable microns (*Zent et al.*, 1993; *Schorghofer and Aharonson*, 2005), this quantity represents a significant fraction of the present atmospheric water abundance. Regions with high dust contents may also incorporate large volumes of adsorbed water which can be gained and lost on diurnal and annual timescales. The ability of regions with high storage capacities to saturate in a given time, and the rate at which they remove water from or add water to the atmosphere, will depend on the diffusion coefficient of the soil. The presence of such large or active subsurface reservoirs will affect the accuracy of models of the Mars water cycle.

GCMs also investigate the water cycle during past climate epochs when surface ice was stable at locations different from those that currently harbor seasonal frosts (*Mellon and Jakosky*, 1993; *Chamberlain and Boynton*, 2007) and when precipitation may have been an active process. (*Haberle et al.*, 2003; *Mischna and Richardson*, 2005). The necessity of incorporating the exchange of subsurface ice, and the changing positions and volumes of stable reservoirs, grows in importance as the time span covered by these models becomes comparable to or exceeds subsurface–atmosphere communication timescales. Employing both the high-fidelity tracking of atmospheric conditions provided by GCMs and the evolution of subsurface reservoirs on long timescales would be limited by the computation speed of the GCM system. One solution employs the GCM to compute a steady climate state (requiring approximately 10 model years) and uses these steady-state values as constraints on physically simpler yet much faster evolution models of water reservoirs such as the polar caps, subsurface ice, and surface ice deposits. These models would be run in series, passing terminal conditions to subsequent stages as input parameters. A coupled model such as this would lead to a better understanding of Mars' long-term water cycle, of regions of surface and subsurface ice stability, of the variation of atmospheric water content over millions of years, and possibly of the formation of climate-controlled terrain such as the polar layered deposits.

Both long- and short-duration models would be improved by the incorporation of spatially varying regolith diffusion coefficients. The demonstrably heterogeneous nature of the martian regolith, as observed by landers and from orbit, implies that the diffusion coefficients of various regions will be different. Even given the result that many loose-particle materials exhibit similar diffusion coefficients (see Chapters 3 and 4 and *Hudson et al.*, 2007; *Hudson and Aharonson*, 2008), the effect of minor variations could accumulate when considered over large spatial or temporal scales. By relating pore size to particle size, and determining the global variation of the latter using measurements obtained from orbit, a rough global map of diffusivity may be constructed. One step of this process, relating observable surface properties to particle size, has been performed for martian fines using reflectance spectra (*Ruff and Christensen*, 2002), and for a broader range of particle sizes using thermal inertia (*Kieffer*, 1976; *Jakoksy and Christensen*, 1986; *Mellon et al.*, 2000). In *Mellon and Phillips* (2001), distinct thermal inertia and albedo units were defined from Thermal Emission Spectrometer measurements. Spatial variations in diffusivity may be assigned based on such surface units if the particle size derived from thermal inertia measurments correlates, even broadly, with diffusivity. Using such a map as a set of input parameters for GCMs, the effect of spatially varying diffusion coefficients on the water cycle could be assessed. Caution must be exercised in relating observable surface properties to pore size, or even in relating particle size to pore size, since the correlations depend on many factors, some of which are still poorly constrained.

1.6.2 Phoenix Scout lander

The Phoenix Scout mission is the first robotic exploration of Mars deliberately targeted at a site expected to possess water in the form of subsurface ice. Landing site selection has been based in part on Mars Odyssey measurements, but also on theoretical predictions of ice table equilibrium depth at high latitudes (*Arvidson et al.*, 2008). These investigations indicate the presence of ice at the landing site within a few decimeters of the surface (*Boynton et al.*, 2006; *Mellon et al.*, 2008).

The Robot Arm on Phoenix will be used to excavate up to half a meter into the surface, exposing and potentially sampling the ice. Analyses of soil thermal conductivity and surface albedo will help further refine subsurface thermal models and predictions of ice table equilibrium. The depth of the observed ice table, if different from the predictions, may be used to refine both the history of the atmospheric vapor content at the site and the diffusion coefficient of the soil. This latter endeavor will also be assisted by observations of soil particle sizes with the Optical Microscope.

By examining the ice table itself, the habit of the subsurface ice, whether it exists as a pore-filling substance or as a massive, regolith-poor layer, will be revealed. This observation directly tests the ice age model of *Schorghofer* (2007) and the models of Chapter 6, which predict massive ice retreat followed by diffusive replenishment, giving a present-day configuration of unsaturated pore-ice near the equilibrium depth. If the filling fraction can be estimated through qualitative measurements of ice mass and regolith porosity, or perhaps correlated with the mechanical strength of the ice-bearing layer, further models of ice growth rates can be assessed. Future missions may glean further information about water in Mars' past by examining subsurface ice distributions with ground-penetrating radar or coring into ice-rich ground; they may also extract isotopes and trapped gases from subsurface ice or polar ice sheets, which are expected to hold a record of the planetary climate, as do ice sheets on Earth (*Murray et al.*, 1972; *Cutts and Lewis*, 1982; *Laskar et al.*, 2002).

1.7 This Work

In Chapter 2, the theory of gaseous diffusion is reviewed, with particular emphasis on diffusion in porous media and the conditions and processes which prevail on the surface of Mars. Chapter 3 covers the materials, methods, results, and interpretations of experiments which simulate buried ice loss into a dry martian atmosphere. The content of these chapters was previously published in *Hudson et al.* (2007).

Chapter 4 extends the investigations of Chapter 3, using the same experimental procedure, to examine enhanced diffusion barriers in the form of salt-encrusted soils, mixtures of sand and dust particles, and fine dust—both compacted and uncompacted. The content of this chapter is published in *Hudson and Aharonson* (2008).

In Chapter 5, the converse process to the ice loss in Chapters 3 and 4 is investigated. Atmospherically derived water vapor is driven down a static, thermally induced vapor density gradient to fill the pore spaces of an initially dry regolith. This represents the first experimental investigation of diffusively derived pore ice under any conditions. The measured ice content profiles are compared to numerical models of vapor diffusion and deposition. The phenomenon of pore space constriction is emphasized, since the experiments indicate that the effect of deposited ice on the diffusion coefficient may not be linear, as assumed in previous models (*Mellon and Jakosky*, 1995). The content of this chapter has been submitted for publication as *Hudson et al.* (2008).

Numerical models investigating the evolution of cryospheric ice over several obliquity cycles are the subject of Chapter 6. These models use Mars' history of orbital parameters and insolation variation to determine how and to what extent parameters such as deep ice content, ice-free surface-layer thickness, and constriction non-linearity affect the distribution of pore-filling ice, which has accumulated since predicted loss events of 300–600 thousand years ago.