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Editorial

Introduction to the 4th Mars Polar Science and Exploration Conference special issue: Five top questions in Mars polar science

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ABSTRACT

As an introduction to this *Icarus* special issue for the 4th Mars Polar Science and Exploration Conference, we discuss five key questions in Mars polar science, gleaned from plenary discussions and presentations held at the conference. These questions highlight major unknowns in the field. (1) What are the physical characteristics of the polar layered deposits (PLD), and how are the different geologic units within, beneath, and surrounding the PLD related? (2) How old are the PLD? And what are their glacial, fluvial, depositional and erosional histories? (3) What are the mass and energy budgets of the PLD, and what processes control these budgets on seasonal and longer timescales? (4) What chronology, compositional variability, and record of climatic change is expressed in the stratigraphy of the PLD? (5) How have volatiles and dust been exchanged between polar and non-polar reservoirs? And how has this exchange affected the past and present distribution of surface and subsurface ice?

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1. Introduction

Given the immense influence of the climate on surface geology, atmospheric evolution, and habitability, understanding Mars' climate history is of crucial importance to the Mars science community. By analogy with terrestrial ice core climate studies, the polar deposits (both the PLD and the residual caps) may contain the most complete existing record of recent climate change on Mars. In effect, the polar deposits could potentially be a Rosetta Stone for the martian climate. In order to decipher this Rosetta Stone, we must understand the geologic history of the polar regions, the ages of the polar deposits, their melting and flow histories, their stratigraphy, and their interactions with the martian climate and with lower latitude ice deposits.

To promote and advance Mars polar science and exploration, the Lunar and Planetary Institute has organized a series of conferences. The Fourth Mars Polar Science and Exploration Conference was held October 2–6, 2006 in Davos, Switzerland (see www.lpi.usra.edu) and was convened by Stephen Clifford (Lunar and Planetary Institute), Walter Amman (Swiss Federal Institute for Snow and Avalanche Research), Kathryn Fishbaugh (International Space Science Institute), David Fisher (Geological Survey of Canada) and James Head III (Brown University). The 108 participants represented a broad range of expertise in terrestrial and martian glacial geology, glaciology, and atmospheric science. Papers contributed in association with the 1st (Camp Allen, TX, in 1998), 2nd (Reykjavik, Iceland, in 2000), and 3rd (Alberta, Canada, in 2003) conferences have also been published in dedicated special issues of *Icarus* (Clifford et al., 2000b, 2001, 2005).

In this introductory paper, we discuss five key questions in current Mars polar science that reflect major gaps in understanding

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of the martian polar regions. We have formulated these questions from presentations and plenary discussions held during the conference. The purpose of this paper is to highlight key unresolved questions and needed observations as a proposed guideline to future Mars polar science and exploration, rather than to provide a comprehensive review of the state of knowledge of the field. However, we do briefly concentrate on advances since the 3rd conference and since the review provided in Clifford et al. (2000a).

2. What are the physical characteristics of the polar layered deposits (PLD) and how are the different geologic units within, beneath, and surrounding the PLD related?

Planum Boreum in the north (Fig. 1) is a broad topographic dome consisting of, in stratigraphic order from top to bottom, the north polar residual ice cap (the portion of the surface which remains relatively bright year-round), the classic north polar layered deposits (NPLD), and layered materials variously called the north polar basal unit or the Scandia region unit and the Rupes Tenuis unit. Planum Australe in the south (Fig. 2) is also a broad topographic dome consisting of the southern residual ice cap (again, the portion of the surface which remains relatively bright year-round) and the classic south polar layered deposits (SPLD). The residual cap and PLD portions of these polar plateaus are among the most recent geologic deposits on Mars.

2.1. How well linked are the north and south PLD, and can we explain the fundamental differences between them?

Both the north and south PLD consist mostly of water ice, are internally layered and have presumably been deposited by similar processes. Commonalities in the history of the north and south PLD may be discernable in their layering, as discussed in Section 5. But the gross similarities appear to end here.

In many of the details, the two polar plateaus on Mars are significantly different from each other. Planum Australe appears to be almost completely covered by dust, while Planum Boreale is mostly covered by the northern residual ice cap (NRC) and by dust in the troughs. The dominant control of high obliquity on the survival and extent of the PLD can be tempered by the presence of such a surficial dust deposit which can insulate against high summer temperatures and inhibit the sublimation of the underlying ice during periods of high obliquity (see Section 6). Indeed, the effects of a dust mantle may explain the older age of the SPLD, which appear to have survived through many high-obliquity cycles (see Section 3). It is not known whether both polar PLDs are in the same state of net mass balance, whether one set of PLD is feeding or has fed lower-latitude ice deposits, or whether the SPLD are currently dormant due to their dust covering.

Currently, the SPLD are partially covered with a semi-permanent layer of CO_2 ice (the southern residual cap, SRC), but it has recently been suggested that this layer forms and disappears in a cyclic manner (Byrne and Zuber, 2006). The long-term stability of this CO_2 ice is unknown, as is how representative its presence has been throughout the history of the SPLD. Has the northern plateau ever been covered by a perennial deposit of CO_2 ice? Richardson and Wilson (2002) suggest that the current lack of permanent CO_2 ice in the north is due to asymmetries in annual mean circulation driven by the elevation difference between the poles. Currently, this circulation pattern favors water ice deposition and retention in the north.

The absolute elevation of the SPLD, as measured by MGS Mars Orbiter Laser Altimeter (MOLA), is ~6.5 km higher than the NPLD (Zuber et al., 1998). Scarps carve into the southern plateau, a few concentric to tangentially oriented troughs and ridges (Ultimi Scopuli), whereas the NPLD display pronounced systems of spiraling troughs and locally complex marginal scarps (Boreales and Gemini Scopuli). Both plateaus have large chasmata, but the origin of these features remains poorly understood, with potential explanations including sublimation, katabatic wind erosion, and melting (Clifford, 1987; Benito et al., 1997; Anguita et al., 2000; Howard, 2000; Kolb and Tanaka, 2001; Fishbaugh and Head, 2002; Kolb and Tanaka, 2006). Why did chasmata form at both plateaus, and did they form at the same time or by similar processes? We briefly discuss the formation of these chasmata in Section 3.

Some of these differences may be explained by the elevation difference between the polar plateaus and differences in their local environments, such as topographic constraints on temperature, atmospheric circulation and local winds, possible volcanic eruptions and/or impact events in the local environment, differences in the grain size of the ice and its change with depth due to different local conditions, differences in the composition and abundance of the local dust sources, and differences in basal topography affecting melting and flow, etc. There have been few comprehensive, comparative studies that address the differences between the north and south PLD.

2.2. What is the nature of the geology beneath the PLD, and how does it influence their form and history?

Geologic mapping indicates that the layered deposits at both poles may be much younger than, and have little relation to older, underlying materials (e.g., Tanaka and Kolb, 2001). An exception may be the north polar basal unit (also called the Scandia region and Rupes Temis units; Tanaka et al., 2008) and divided into various subunits; Tanaka and Bourke, 2007], which stratigraphically lies between the Hesperian Vastitas Borealis materials and the NPLD. The origin of this unit, which is the main source of the north polar sand sea, is still under debate, with hypotheses ranging from alternation of sand deposition (Byrne and Murray, 2002; Edgett et al., 2003; Fishbaugh and Head, 2005) with ice deposition (Byrne et al., 2007; Russell et al., 2007) to layers created from reworking of Vastitas Borealis and Scandia region materials (Tanaka, 2005; Tanaka and Bourke, 2007; Tanaka et al., 2008). The Byrne et al. (2007) and Russell et al. (2007) hypothesis indicates an intimate relationship between the PLD and basal unit. In this hypothesis, wherein the PLD began to build up when the basal unit sand supply (or northward sand migration) ceased, allowing buildup of only the ice-rich layers, the main difference between the basal unit and the overlying PLD is sand content. In the south, Planum Australe does not appear to have any counterpart to the northern basal unit lying stratigraphically beneath the SPLD.

The Mars community has raised significant questions about the nature of the geology under the polar plateaus and about the possible relationship of that geology to the NPLD and SPLD. The basal topography is presently being mapped by subsurface radar from the Mars Express (MEx) Mars Advanced Radar for Subsurface and Ionosphere Sounding (MARSIS) instrument (Plaut et al., 2007) and the Mars Reconnaissance Orbiter (MRO) Shallow Subsurface Radar (SHARAD) (Phillips et al., 2008). The radar mapping may constrain the upper limit of isostatic depression in the SPLD. The radar mapping of the internal layering may also help reveal clues to the geology under the ice and to subsurface thermal anomalies beneath the polar caps by noting disturbances in layer continuity, and whether, in part, subsurface topography does control the location of major reentrant features, like Chasma Boreale and Chasma Australe.

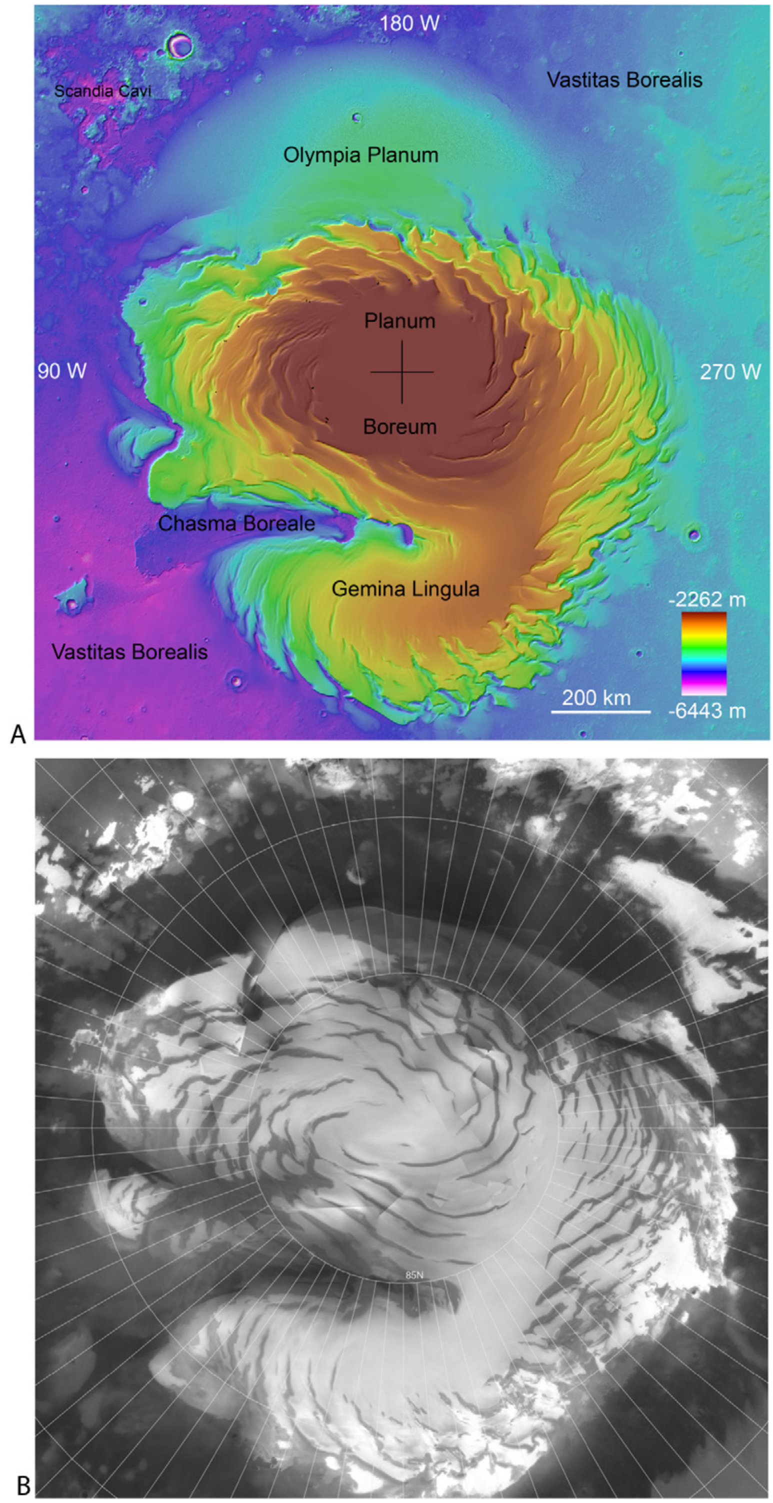

Fig. 1. (A) MOLA shaded relief and topography (512 pix/deg) of the north polar region with major geographic features labeled. (B) Portion of MOC image mosaic of MC01 Quadrangle (NASA/JPL/Malin Space Science Systems), same area as in (A). From: http://www.msss.com/mars_images/moc/moc_atlas.

3. How old are the polar layered deposits? And what are their glacial, fluvial, depositional and erosional histories?

Determining how long the PLD and residual caps have existed is intimately linked to the interpretation of the exposed stratigraphy, to whether or not ice flow has been significant, to long-term variations in the geographical location and extent of the PLD, and to the D/H ratio of the ice.

Using crater ages and assuming insignificant ice flow, Cutts et al. (1976) estimated that the NPLD are about 10 Myr old at most, and Tanaka (2005) more recently halved this estimate. Now, Tanaka et al. (2008) suggest that Olympia Planum is made up of NPLD, bringing the age estimate up to ∼1 Byr. Similarly, using crater-derived resurfacing rates, Plaut et al. (1988) estimated the SPLD to be ∼10 Myr, but Koutnik et al. (2002) have revised this to ∼30–100 Myr. It is important to note that crater-counting

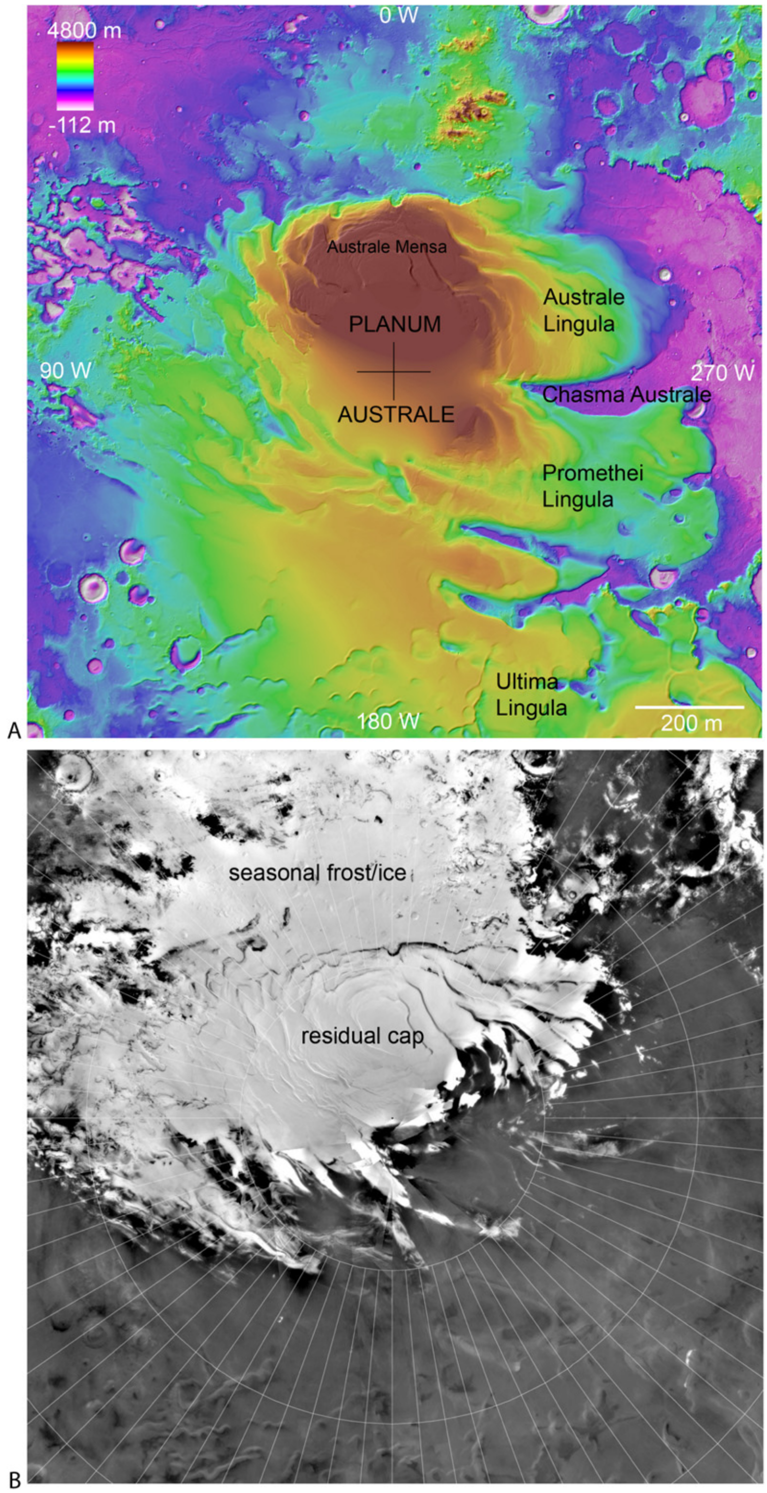

Fig. 2. (A) MOLA shaded relief and topography (512 pix/deg) of the south polar region with major geographic features labeled. (B) Portion of MOC image mosaic of MC30 Quadrangle (NASA/JPL/Malin Space Science Systems), same area as in (A). From: http://www.msss.com/mars_images/moc/moc_atlas.

ages are merely surface exposure ages and tell us little about the age of the entire stack of PLD. The presence of a dust lag may be crucially effective in protecting the SPLD from sublimation, resulting in an orders of magnitude older surface crater age than for the north. MARSIS data have recently indicated that the SPLD are rich with internal radar layering (Picardi et al., 2005; Plaut et al., 2007), possibly indicating the presence of similar, older protective lag deposits that could imply an even older age for the entire SPLD than suggested by the surface crater counting. One must also allow for possible lag deposits with similar effect in the NPLD, possibly corresponding to internal radar layering or even to some of the visible layers. Processes such as ice accumulation, ice ablation, viscous relaxation, ice flow, and aeolian reworking can also complicate age determinations by altering the cratering record on the exposed ice surface. Due to the influence of these modification processes on the polar deposits, the surface exposure age and therefore the age of internal layers are poorly constrained from the cratering record alone.

3.1. Have the polar layered deposits ever flowed and, if so, at what rates and under what conditions, and what is the current role of flow in controlling the shape of the deposits?

Other aspects of polar history may prove useful in constraining the age of the PLD. Most authors agree that large-scale ice flow within the PLD is at present extremely slow (to the point of being negligible) (Fisher, 2000; Greve et al., 2003; Hvidberg, 2003; Fishbaugh and Hvidberg, 2006). However, the presence of putative deformation features in both the south and north PLD (Herkenhoff et al., 2003; Milkovich and Head, 2006) imply disturbances resulting from a more dynamic flow field in the past. Indeed, several studies have concluded that under higher obliquity conditions, ice flow and ice relaxation may have been considerably faster than at present (Greve et al., 2003; Hvidberg, 2003; Pathare and Paige, 2005; Pathare et al., 2006; Winebrenner et al., 2006). If putative flow features stand the test of image photoclinometry and MRO High Resolution Imaging Science Experiment (HiRISE) and Context Camera (CTX) stereo analysis, and if studies of PLD-wide layer structure (via imaging and radar data) indicate significant evidence of effects of flow on that structure, then we must conclude that the PLD are old enough to have experienced at least one period when the mean obliquity was substantially higher that at present to allow for warmer ice and faster flow.

3.2. What can stratigraphy tell us about the age of the polar deposits?

In Section 5, we describe the stratigraphy of the PLD. Obviously, that stratigraphy must contain time information and must reveal something about the age of the PLD, yet the relationship is still unclear. It has been assumed since Mariner 9 that the layering of the PLD reflects changes in mass balance resulting from cyclical changes in orbital parameters (Cutts and Lewis, 1982). Several studies have attempted to relate brightness profiles in MOC images with orbital parameters using different methods (Laskar et al., 2002; Milkovich and Head, 2005). Such studies allow dating of layers by inferring a timescale from calculated changes in orbital parameters with respect to time (Laskar et al., 2004). But if inherent layer brightness is masked by surficial dust and frost/ice deposition (e.g., Milkovich and Head, 2006; Herkenhoff et al., 2007), then tying such profiles directly to changing orbital parameters may not be the most reliable method for dating individual layers; thus, their timescale is uncertain. However, as described in Section 5, the depositional and erosional history of the PLD can be determined from stratigraphic analysis, even if not directly related to time, by determining whether layer deposition was areally uniform (layer continuity), by inferring mass balance patterns through time (via layer thickness), and by searching for large and small scale unconformities that indicate periods of minimal deposition or net erosion. We can at least state that if layer deposition is indeed controlled for the most part by changes in insolation (orbital parameters), then the PLD extend back far enough in time to allow insolation cycles to create the existing number of layers and (erosional) unconformities, perhaps experiencing many tens to hundreds of cycles.

Obviously, the age of materials underlying the PLD puts a maximum constraint on PLD age. While much of the north polar basal unit is buried beneath the PLD, making crater dating of its surface difficult, its presence at least implies that there was a time in the past when deposition of the sand-rich basal unit material was favored over thick accumulations of ice-rich, PLD material, at least in the north (possibly during a higher average obliquity?). Clearly, dating of the PLD is crucial to understanding the history of the planet, clues to which may be preserved in the polar ice. But much more work remains to be done before that chronology is defined and understood.

3.3. Is there evidence of past or present melting, and how does this relate to age?

While still a controversial topic (e.g., Tanaka and Bourke, 2007), several authors have suggested that the formation of Chasma Boreale and Chasma Australe may have been initiated by geothermal melting and fluvial (not necessarily catastrophic) discharge (Clifford, 1987; Benito et al., 1997; Anguita et al., 2000; Fishbaugh and Head, 2002). The putative Chasma Boreale melting event has also been proposed as a possible source of water for the creation of gypsum within the Amazonian-aged north polar sand sea (Fishbaugh et al., 2007b). It has been suggested that large-scale retreat of the NPLD (Fishbaugh and Head, 2000) exposed the basal unit to reworking into Amazonian-aged dunes (Byrne and Murray, 2002; Edgett et al., 2003; Fishbaugh and Head, 2005). Thus, if melting played a part in forming gypsum within these dunes, it must have occurred after the putative retreat of the NPLD. Did orbital parameter-induced climate change cause melting and retreat or were such events caused by some other endogenic trigger (e.g., a sub-ice volcanic eruption)? If the latter is true, then age constraints may be provided by such indirect means as determining when, in the general history of Mars, such an endogenic event could have taken place.

3.4. What can hydrogen isotopes tell us about the age of the polar deposits?

Hydrogen isotope ratios (D/H) of the water derived from the PLDs, residual caps, the shallow subsurface, and lower latitude ice deposits would place constraints on the age and/or accessibility of the water sources for a growing PLD or residual cap and on water sinks for a shrinking PLD or residual cap. The stable isotopic ratios for all of these ice reservoirs are coupled since the PLD and/or residual caps may be the ultimate source for lower latitude ice, via sublimation and re-deposition from the atmosphere (see Section 6).

A very old PLD age would imply that the PLD are relatively stable as a whole, meaning that PLD water cannot migrate quickly or often to other parts of Mars to make lower latitude ice deposits. In such a case, we must envisage another significant water source for that function. If there are no other significant water reservoirs, then the bulk average D/H of the (very old) PLD is likely to be much lower than that of a young PLD (Fisher, 2007). The reason for this lower ratio is that little of the ice has been sublimated into the atmosphere and hence little hydrogen has been lost by atmospheric escape relative to deuterium. If the PLD are young, $\sim 5 \times 10^6$ years old at the bed, then, by implication, water might be able to come and go on shorter timescales. The D/H ratio in this case would be higher than for older PLD. However, if the ice cap is young AND still has a low bulk D/H, then there must be much more accessible water than found in the PLDs, which helps to buffer escape of PLD-only derived hydrogen.

The Novak et al. (2005) D/H seasonal measurements tend toward a low minimum D/H that is coincident with the maximum precipitable atmospheric water abundance. These observations tend to support an old PLD, if the PLD are the main source of that atmospheric water. What is needed to resolve this issue is more seasonal atmospheric D/H data and a history of D/H from the PLD. Sampling the stratigraphy of the PLD—by drilling, a cryobot, or traversing down a trough wall with a rover—and measuring the water isotopes is an important goal. Any relationship between climate cycles and the orbital cycles should show up in the layers of the ice cap as variations in water ice accumulation, dust, and stable isotopes (D/H). Additionally, solar variability (solar wind) should imprint on the D/H ratio, with the same time variations observed in terrestrial paleo records. The sampled stratigraphy of the PLD should be datable using the various orbital cycles in the climate and the solar wind. Atmospheric D/H ratios can in principal be obtained remotely. Novak et al. (2005) have shown this can be done from Earth, and deploying a version of their spectrometer on an orbiter could prove extremely useful.

4. What are the mass and energy budgets of the PLD, and what processes control these budgets on seasonal and longer timescales?

The mass balance of an ice mass is the annual difference between accumulation and ablation. The annual mass balance of the martian polar deposits consists of seasonal accumulation and sublimation of H_2O ice and CO_2 ice and deposition and removal of dust. It is important to remember, throughout this discussion, that the seasonal and residual caps may be the only polar deposits currently in direct communication with the atmosphere. Dust blankets much of the exposed portions of both the north (in troughs) and south (in troughs and on the surface) PLD.

To understand the current and past mass budgets of the polar deposits, the physical processes that contribute to accumulation and ablation must be understood. Orbiters and landers provide critical data to study the modern day processes, and although much is now known about the behavior of the seasonal CO_2 cap and atmospheric H_2O, the physical processes are still not well understood. Variations in orbital parameters, in particular the obliquity, exert a strong control on variations of the mass and energy budgets of the polar deposits through their effect on the distribution of solar insolation. The role of dust on climate variations and its effect on volatile cycles has yet to be fully explored, in particular how a layer of dust could reduce the exchange of volatiles between the atmosphere and surface ice deposits.

Developing a better understanding of present day processes may suggest what conditions were required for ice to accumulate at the poles in the past, constrain the age of the PLDs (see also Section 2), and, together with the geologic data (see Section 2), a history of ice formation in the polar regions may emerge. This question encompasses a large number of uncertainties due to the fact that direct measurements of mass balance have never been made.

4.1. How does the current radiation budget vary with season, and how is it affected by the presence of ice and dust in the atmosphere?

Present-day Mars is a radiation-dominated environment. Obliquity (see Section 6) and eccentricity control seasonal variations in the radiation budget, and the atmospheric opacity has a large control on surface temperature. Mars has an elliptical orbit around the Sun, with an eccentricity of 0.093; therefore, seasons are more exaggerated than for Earth whose orbital eccentricity is less. Currently, the shorter southern hemisphere summer occurs near perihelion, resulting in short, hot southern summers and long, but relatively cool northern summers. Changes in accumulation and ablation due to obliquity and eccentricity changes lead to asymmetries in mass balances of the north and south PLD and also to the growth and decay of mid-latitude ice deposits (see also Section 6).

Changes in atmospheric opacity are due primarily to dust loading. Seasonally, dust deposition at the poles lowers the ice-surface albedo, increasing the amount of radiation absorbed at the surface. In addition, CO_2 deposition during the fall and winter alters the radiation budget by increasing the albedo and increasing the amount of reflected radiation. CO_2 deposition and CO_2 ice clouds may affect infrared (IR) emission which could explain anomalous Viking Infrared Thermal Mapper (IRTM) brightness temperature measurements during polar winter (Forget and Pierrehumbert, 1997). The martian atmosphere is almost entirely composed of CO_2, which is a strong absorber of IR radiation. The full effects of CO_2 deposition and atmospheric dust needs to be more thoroughly characterized and quantified.

4.2. What is the current mass balance (mechanisms, rates, temporal and spatial variability) of CO_2, H_2O and dust?

The seasonal CO_2 cycle, including formation of the seasonal caps, has been extensively studied (recent references include, e.g., James and Cantor, 2001; Kieffer and Titus, 2001; Smith et al., 2001; Piquex et al., 2003; Douté et al., 2006). Each year, seasonal insolation variations result in about 25% of the martian atmosphere alternatively condensing and sublimating from the poles. In the north, no CO_2 survives the summer; however, in the south, a semi-stable layer of CO_2 survives throughout the year as the SRC (Leighton and Murray, 1966; Tillman et al., 1993; Forget et al., 1998). This seasonal cycling of CO_2 is driven primarily by insolation variations, as first predicted by Leighton and Murray (1966). The present-day seasonal cycling of CO_2 has been investigated using energy balance and General Circulation Models and by tracking seasonal cap changes in: surface neutron flux (GRS; e.g., Feldman et al., 2003), surface topography (MOLA; e.g., Smith et al., 2001; Aharonson et al., 2004), thermal inertia (TES; e.g., Kieffer and Titus, 2001), and ice grain size inferred from imaging spectrometer observations (OMEGA; e.g., Langevin et al., 2005). However, while the overall behavior and composition of the seasonal cap is well-known, the thickness measurements have a precision close to the actual seasonal cap thickness, so that its mass balance has not yet been determined.

Many mechanisms contribute to form, erode and modify the residual caps and PLD. Surface deposition of H_2O and dust occurs together with seasonal deposition of CO_2. Atmospheric dust raised from the surface by dust storms is thought to act as condensation nuclei for the water vapor and atmospheric CO_2 (Kieffer and Titus, 2001). In autumn, H_2O and CO_2 condense on the dust particles and precipitate to the surface (reviewed in Clifford et al., 2000a). H_2O and CO_2 frost may also condense directly on the surface or in pores in the near-subsurface depending on temperature and air humidity (Schorghofer and Aharonson, 2005). In the south, the formation of the seasonal cap is thought to occur primarily by precipitation, whereas, in the north, deposition may occur primarily by direct condensation of vapor (Colaprete et al., 2005). In the spring, sublimation recycles the seasonal CO_2 frost back into the atmosphere, although a residual deposit of dust and H_2O may remain on the surface. The residual CO_2 layer covering the SPLD may reflect decadal or centennial variations in the CO_2 cycle (Byrne and Ingersoll, 2003), which may affect the H_2O cycle. Winds may erode the surface, redistribute H_2O on the ground and recycle vapor back to the atmosphere. Diurnal and annual temperature variations in the near-surface layers could redistribute ice there. These processes would smooth the surface, compact the material, and round the grains, which appears consistent with the interpretation of OMEGA observations (Langevin et al., 2005). Currently, the effects of seasonal and interannual variations may not affect both polar deposits in the same way. For example, dust covering much of the SPLD surface could serve as protection against insolation changes.

Richardson and Wilson (2002) suggest that the difference in elevation between the two caps influences H_2O deposition, with the NPLD being favored under current conditions. Water ice that sublimates as the seasonal CO_2 retreats from the north pole is hypothesized to be driven poleward by baroclinic eddies ("vacuuming effect"), resulting in accumulation at the highest latitudes (Houben et al., 1997; Bass et al., 2000). Deposition of dust and ice may also be influenced by the effect of mid-latitude topography on atmospheric circulation. In particular, Colaprete et al. (2005) attribute the offset of the southern residual cap from the pole to this offset. At each pole, the geographic variations in deposition are likely controlled by local environmental conditions such as topography, winds, albedo, etc.

Estimates of recent absolute accumulation rates of the NPLD, based on resurfacing rates inferred from crater counting and on tying layer brightness patterns to orbital cycles, range from ~0.01 to ~1.2 mm/martian-yr (Plaut et al., 1988; Herkenhoff and Plaut, 1999; Laskar et al., 2002; Milkovich and Head, 2005). Spatial and temporal variations in the mass balance of the NPLD have been inferred from thicknesses of correlated layer sets in MOC images (Fishbaugh and Hvidberg, 2006). Brightness may also indicate where deposition or erosion is actively occurring. While Hale et al. (2005) find that the cap margins tend to brighten more consistently from year to year than the interior areas based on a 3 year MOC study, the study by Bass et al. (2000) using Viking images did not exhibit such a trend. Albedo variations have been summarized by both Hale et al. (2005) and Cantor et al. (2002) as being complex, though recent MRO/MEx observations are hinting at consistent albedo patterns on the NRC (Calvin et al., 2007). It may also be possible to infer from where sublimation is occurring from increases in atmospheric water vapor, such as the high values observed over Olympia Planum at 75–80° N, 210–24° E (Melchiorri et al., 2006). Direct measurements of mass balance have not yet been attempted; thus, it is only loosely constrained by inference and modeling. Measurements of annual or even decadal layer thicknesses and a reliable dating system can only be accomplished with analysis of a (several) sample core(s).

5. What chronology, compositional variability, and record of climatic change is expressed in the stratigraphy of the polar deposits?

Climate change may be expressed in the polar stratigraphy by variations in layer composition, texture, and thickness, but the implications of these differences for the climatic conditions that produced them is poorly understood; nor do we understand the chronology the layers reflect.

5.1. How can the internal layers be dated (relatively and absolutely), and what portion of Mars' history do these layers represent?

The impact cratering record currently provides the primary source of age information for Mars, and this record provides a (possibly inaccurate) surface exposure age for the polar deposits, as described in Section 3. In an attempt to date the individual polar layers, Laskar et al. (2002) matched changes in the apparent brightness of individual NPLD layers in a MOC NA image to calculated obliquity variations, constrained by the young surface age implied by crater counts. However, given the previously discussed uncertainties associated with crater-dating the surface, it is possible that the surface is much older than implied by the small number of identifiable craters. If so, it may be necessary to match layer brightness profiles to other portions of the obliquity history (Pathare and Murray, 2006). Uncertainties in crater dating are not the only problem to plague attempts to establish a chronology for the layers. If apparent layer brightness does not match inherent layer brightness, then, as described below, characterizing layers by their brightness alone may be unreliable. Additionally, we do not know if the layers are a continuous record through time or if layers have been lost during episodes of erosion. Thus, we need to better understand the detailed layer stratigraphy, as described below.

5.2. What is the H_2O, CO_2, and dust content of the residual caps and PLD, and how does composition correlate with layer stratigraphy?

Various observations indicate that the north polar residual ice cap (NRC) is composed largely of water ice (Kieffer et al., 1976; Kieffer and Titus, 2001; Langevin et al., 2005). However, the albedo of the NRC surface is variable, with values consistently slightly less than that for pure water ice (Bass et al., 2000; James and Cantor, 2001; Kieffer and Titus, 2001; Malin and Edgett, 2001; Hale et al., 2005). Such albedo variations can be attributed to the presence of dust impurities and to changing grain size as the seasonal frost cap sublimates (Warren and Wiscombe, 1980; Kieffer, 1990; Langevin et al., 2005). Although it can be difficult to estimate the dust/ice ratio from albedo alone, compositional modeling using OMEGA data indicates a maximum NRC dust content of about 6% (Langevin et al., 2005). The southern residual ice cap (SRC), by contrast, is composed largely of CO_2 ice (Kieffer, 1979), but many observations also indicate an additional component of water ice (Barker et al., 1970) in areas not covered by CO_2 and beneath the CO_2 (Byrne and Ingersoll, 2003; Bibring et al., 2004; Hansen et al., 2005). This CO_2 ice cap may not be a permanent feature of the south polar deposits (Jakosky et al., 2005), but rather may be built-up (at either pole), during periods when major dust storms coincide with polar summer and sublimated at other times (Byrne and Zuber, 2006). Thus, the water ice component of the SRC may simply be the southern counterpart to the NRC peeking through the temporary CO_2 ice covering.

Viking color data of NPLD exposed in trough walls are consistent with the presence of at least a partial bright, red dust component (Thomas and Weitz, 1989; Herkenhoff and Murray, 1990). Until MRO's Compact Reconnaissance Imaging Spectrometer (CRISM) definitively identified water ice as a component of the NPLD, the presence of water ice was deduced by the observed relationship between surface temperature and atmospheric water content (Farmer et al., 1976; Kieffer et al., 1976). The CRISM data exhibit variations in the depth of the 1.5 μm water absorption band within the NPLD stratigraphy exposed in Chasma Boreale scarps; such variations can be attributed to both H_2O ice grain size changes and to variations in the H_2O ice/dust ratio (Murchie et al., 2007). Recent observations of the NPLD by MRO HiRISE have shown in detail what has long been suspected: that the albedo of the PLD exposed in trough walls and scarps is at least partially dependent upon lighting conditions and upon younger, surficial deposits of dust, ice, and frost controlled by local and small-scale topography (Herkenhoff et al., 2007). Since imagers and imaging spectrometers only observe the upper few microns of a material, the ice/dust ratio of the rest of both PLD must be inferred from these observations and by other means.

The MEx MARSIS radar has been able to penetrate to the base of the NPLD (Picardi et al., 2005) and SPLD (Plaut et al., 2007) and the MRO SHARAD radar to the base of the NPLD (Phillips et al., 2008). The radar observations indicate that both PLD must be composed almost completely of ice, with dust making up only a few percent of the volume. Compositional observations alone are thus consistent with the dusty, icy NRC simply being the most recent layer of the dusty, icy NPLD, and perhaps the same is true of the southern water ice hidden beneath the CO_2 ice.

Continued observations by MEx, Mars Odyssey, and MRO will likely lead to more revelations about the composition of the N/SRC and the N/SPLD, especially with synergistic observations between instruments. Areas of the exposed PLD with the least dust/frost/ice cover give the best estimates of ice/dust ratio from orbit. However, all imaging and spectral observations are limited to the upper few microns of the surface. To gain the most reliable estimates of volume percentages of ice/dust in the N/SPLD (heretofore only inferred from radar data), a landed mission may be necessary. Such a mission would need to carry instruments which could scrape away the younger, overlying deposits or would need to be able to drill through them. Knowledge of the volumetric dust/ice ratios of the N/SPLD and N/SRC would lead to a better understanding of how dust and water cycles lead to deposition at the poles. Knowledge of the volumetric ice/dust ratios of individual layers of the N/SPLD is important for linking layer formation to the past depositional environment (e.g., atmospheric conditions).

5.3. How is the stratigraphy of the PLD related to cyclic variations in insolation and in the global cycles of H_2O, CO_2, and dust?

Variations in obliquity may have a dramatic effect on the stratigraphy of the PLDs (e.g., Levrard et al., 2007) by the associated effects on deposition and sublimation rates of H_2O and by the affected frequency and strength of dust storms. But exactly how obliquity might affect layer thickness and, indirectly, layer composition, is unknown. Without any direct observations of annual layering within, or current mass balances (ice and dust) of the residual caps and PLDs (see Section 4), we also can have no way of knowing exactly how seasonal cycles in CO_2, H_2O, and dust have affected the stratigraphy.

5.4. How is the stratigraphy of the PLD related to episodic events such as impacts, volcanic eruptions, global dust storms, and melting?

As discussed below, the large-scale structure of the stratigraphy, wherein layers can be followed along and across one trough (Fenton and Herkenhoff, 2000; Malin and Edgett, 2001), between adjacent troughs (Kolb and Tanaka, 2001), and even between widely separated troughs (Milkovich and Head, 2005; Fishbaugh and Hvidberg, 2006; Milkovich and Plaut, 2007), suggests that it is related to global climate changes. Episodic events such as impacts and melting would mainly influence the layers on a local scale. Volcanic eruptions could affect the PLD directly by atmospheric deposition of impurities and ashes, which would probably also have large local variations due to wind patterns. Volcanism may also have an indirect and large scale effect by inducing climate change. Since the age of the PLD is unknown, we cannot know whether large scale eruptions (e.g., at Alba Patera), small, local eruptions (Hodges and Moore, 1994; Garvin et al., 2000), and/or deposition of ejecta from large impacts (Wrobel and Schultz, 2004) may have affected the PLD. However, volcanic and impact events during the lifetime of the PLD are probably only very rare and cannot be responsible for the bulk of impurities in the ice. The large dust event that occurred during July, 2007 and was carefully monitored by various instruments on MRO could provide a unique opportunity to observe the effects of such events on the SPLD.

5.5. What is the range in thickness, continuity, and extent of layers throughout both of the PLD? Do any stratigraphic features (such as major unconformities) in the north and south correspond to the same event?

The discovery of the distinctive, knobby "marker bed" in three different MOC images along one trough (Malin and Edgett, 2001) has encouraged attempts at correlation from one location to another in the NPLD of individual layers (Fenton and Herkenhoff, 2000; Kolb and Tanaka, 2001; Malin and Edgett, 2001; Fishbaugh and Hvidberg, 2006) and even of layer brightness profiles (Milkovich and Head, 2005; Milkovich and Plaut, 2007). Further analysis of NPLD layer sequences in HiRISE images (resolution up to ~30 cm/pix) is preliminarily revealing the existence of more than one layer similar to the distinct "marker bed" (Fishbaugh et al., 2007a). At Mars Odyssey Thermal Emission Imaging Spectrometer (THEMIS) VIS resolution (17 m/pix), sequences of layers correlate across wide areas of the SPLD (Milkovich and Plaut, 2007), though correlations with higher-resolution data sets have only been attempted over smaller areas thus far (Byrne and Ivanov, 2004). If layers within the SPLD and NPLD have so far only been correlated at THEMIS and MOC scales (respectively), will they be correlateable at smaller scales? If not, what does that imply about the controls on layering at each pole?

Radar data are ideally suited to assessing the continuity of PLD layers below the elevations of trough floors and between trough walls. Phillips et al. (2008) have identified strong SHARAD reflections within the upper 600 m of the NPLD, but the data exhibit packages of well-defined reflections between 600 and 2000 m depth separated by radar dark packages. Some of the decrease in well-defined layering could possibly be attributed to potentially pervasive fracturing in this part of the NPLD as inferred from HiRISE observations (Byrne et al., 2007). One must keep in mind that the relationship between visible layering and radar layering is not immediately clear or apparent, though preliminary attempts have been made to relate layering evident in THEMIS images to MARSIS layering in the SLPD (Milkovich and Plaut, 2007). To better connect radar and visible layering, tie points between them are needed. Precise elevations of individual radar and visible layers may provide such tie points, and could be attained through high resolution HiRISE and CTX stereo DTMs and with future altimetric data having higher-spatial resolution and vertical accuracy than MOLA. Observations of bore hole stratigraphy would provide even better tie points.

While the above discussion indicates that layers within the upper several hundred meters of the PLD exhibit large-scale continuity, layers near the margin present a different story. Near the NPLD margins, angular unconformities and erosion are more common (Howard et al., 1982; Milkovich and Head, 2005; Tanaka, 2005; Fishbaugh and Hvidberg, 2006), and localized examples also occur within the SPLD (Murray et al., 2001). Analysis of HiRISE and CTX stereo anaglyphs and DTMs will certainly aid in the identification of true angular unconformities (vs "optical illusions" created by complex topography) and in the understanding of the stratigraphic relationships associated with them. Angular unconformities constitute an important part of PLD history because they can be created by retreat and advance of the PLD margin, by erosion and later unconformable deposition, by changes in mass balance patterns, and even by tectonic disturbance of the stratigraphy.

Unconformities which express themselves as wide-scale unconformable deposition on previous layers, rather than as angular, can only be recognized through PLD-wide mapping. Fishbaugh and Hvidberg (2006) have identified one such unconformity through the differing mass balance patterns between two correlatable NPLD layer sets. (Tanaka et al., 2008) separate the classic PLD into two, layered geologic units separated by an unconformable contact: Planum Boreum 1 (forming the bulk of the PLD) and Planum Boreum 2 (~150 thick and, in some places, draping over Planum Boreum 2 layers on north-facing trough walls). (Tanaka et al., 2008) have also noted a generally unconformable contact between the NRC and the underlying NPLD. Is the NRC simply the topmost layer of an NPLD which contains several internal unconformities within it, or, put another way, is the NPLD simply built of many past residual caps?

These types of PLD-wide unconformities may represent the best chance of correlating the stratigraphy within the NPLD to that within the SPLD; the discovery of such unconformities at both poles would strongly indicate that global climate has affected deposition at the poles in a similar manner at various times in the past. Indeed, Kolb and Tanaka (2006) have mapped a major unconformity between two of their mapped geologic units within the SPLD. Ongoing geologic mapping by Kolb and Tanaka (2007) and by

(Tanaka et al., 2008) of both PLD indicate additional stratigraphic complexities that may correlate from one pole to the other.

Since MOC has shown layering down to the limit of its resolution in both PLD, it has been expected that HiRISE will do the same and that with future, higher-resolution cameras, thinner and thinner layers would become visible. Surprisingly, preliminary analysis suggests that HiRISE (∼30 cm/pix, maximum) has revealed the thinnest layers visible from orbit (∼30 cm–1 m); the presence of younger, superposed dust and frost/ice deposits and slumping of this material and possibly of the layers themselves obscures any thinner layering that may be present (Herkenhoff et al., 2007). A shallow drill is needed if the community is ever to observe PLD layering at the scale of annual and decadal deposition.

6. How have volatiles and dust been exchanged between polar and non-polar reservoirs? And how has this exchange affected the past and present distribution of surface and subsurface ice?

The waxing and waning of polar deposits was initially detected by Earth-based telescopic observations (e.g., Herschel, 1784). The time scales over which we have been able to observe and characterize quantitatively the movement of volatiles and dust between polar and non-polar reservoirs is miniscule, however, compared to the length of the geological history of Mars. Furthermore, recent assessments of the historical variations in spin-axis and orbital parameters of Mars (Laskar et al., 2004) predict changes many times larger than those characterizing the recent history of Mars. Thus, we should be acutely aware that past climate conditions on Mars may be completely different from what we see today. Indeed, in terms of spin-axis obliquity alone, we are experiencing an anomalous period relative to the last 10–20 million years, and over the geological history of Mars, obliquity may have exceeded 80°! Such changes would have a profound influence on the location and stability of volatile and dust reservoirs, and even on the very existence of polar caps, during the geological history of Mars. The challenges are (1) to characterize and understand the volatile exchange system under current conditions, (2) to assess the presence and distribution of dust and volatiles as seen in the geological record, and (3) to use atmospheric general circulation models (GCMs) to understand the ancient climate conditions that might explain the geological observations. Progress is being made in all three areas, but many outstanding questions remain.

6.1. What is the current distribution of exchangeable reservoirs of volatiles and dust?

Consideration of the water frost point temperature of the martian atmosphere (typically ∼198 K, the temperature at which the concentration of atmospheric water vapor reaches saturation and below which water ice will condense), and the present range of mean annual surface temperatures [typically ∼150 K at the poles (the CO_2 frost point temperature) to ∼220 K at the equator], suggests that the occurrence of ground ice is generally restricted to latitudes poleward of ∼40° (Fanale, 1976; Farmer and Doms, 1979; Mellon and Jakosky, 1993). At lower latitudes, mean annual surface temperatures are generally high enough, and the concentration of atmospheric water vapor low enough, that near-surface ground ice is unstable. This stability assessment assumes that the gaseous permeability of the regolith is sufficiently high that it will permit water vapor to freely migrate between the atmosphere and subsurface in response to diurnal and seasonal temperature changes; in other words, it assumes that the ground ice is in thermal and diffusive equilibrium with the atmosphere (Zent et al., 1986; Mellon and Jakosky, 1993; Richardson and Wilson, 2002).

These theoretical expectations appear consistent with measurements of the near-surface (top ∼0.5 m) regolith hydrogen abundance made by the Mars Odyssey Gamma Ray Spectrometer (GRS) (Boynton et al., 2002; Feldman et al., 2002). The GRS data are usually interpreted in the context of a simple 2-layer model, wherein ice-rich (>60% by volume) regolith is assumed to underlie a desiccated layer of variable thickness. According to this model, the depth to ground ice varies from ∼13 cm near the poles to ∼50 cm at 40°–50° latitude (Boynton et al., 2002), an increase in depth believed to reflect the greater magnitude and duration of temperatures above the frost point that occurs at lower latitudes. However, recent work has refined previous theoretical models of ground ice distribution by taking into account the high thermal conductivity of ice-cemented soil, bringing the ice table depth about five times shallower than previously predicted, ranging from a few millimeters to a few meters, with a typical value of a few centimeters (Mellon et al., 2004).

The latest calculation of ice table depth seem inconsistent with those modeled using GRS data. It is the thermal wave depth that has the largest control on ice distribution, and this depth is largely dependent on soil properties. The presence of subsurface heterogeneities (e.g., rocks, sand lenses, etc.) can create large local variations in this depth (Sizemore and Mellon, 2006). With a surface resolution of $\sim 3 \times 10^5$ km^2, the hydrogen abundances determined from the GRS data are averaged over areas that are many orders of magnitude larger than the scale of variability observed in the physical and thermal properties of the planet's surface (Malin and Edgett, 2001; Christensen et al., 2003), possibly creating inaccurate results of absolute ice table depth on small areal scales from GRS-based models. This is especially true at mid- to low-latitudes, where the variable, but generally low, abundance of near-surface hydrogen can be interpreted in several ways: (1) it may reflect a region that is uniformly devoid of near-surface ice (at least, within the top meter) but possesses large-scale variations in the abundance of hydrated minerals (e.g., Squyres et al., 2004a, 2004b; Bibring et al., 2005); (2) it may consist of broad areas where the top meter of the regolith is ice-free, but where the diffusion-limiting properties of the local regolith have permitted shallow ice (emplaced under more favorable conditions that may have existed earlier in the planet's history; Smoluchowski, 1968; Fanale et al., 1986; Clifford, 1993) to survive over smaller regions; or (3) it may be attributable to a combination of both explanations.

A number of questions remain both about ice in the circumpolar region and about the characteristics of the regolith overlying the ground ice. Comprehensive mapping is critical to understand the nature and formation of ground ice. We need to know the modern distribution of high-latitude ground ice (in three dimensions), how it varies over different time scales (from seasonal to billions of years), and what causes these variations. A technique to map the depth of the ground ice deposits is needed here.

The overall behavior of the planet's global water cycle is determined by the time-varying influences of local and global sources and sinks, and the general circulation of the atmosphere (e.g., Jakosky et al., 1997; Richardson and Wilson, 2002). Atmospheric water vapor on Mars exhibits strong seasonal variations associated with its saturation vapor pressure (limited by temperature) and the condensation and sublimation of water from surface and near-subsurface reservoirs, such as the PLD, ground ice, and adsorbed water. The maximum depth to which exchange between the atmosphere and regolith can occur is limited to the seasonally active layer (top ∼2–3 m), where the amplitude of the seasonal thermal wave decays to $1/e$.

The reservoirs of water that exert the greatest influence on atmospheric water vapor content are the residual ice caps and PLD. Temperature changes, in response to seasonal and climatic variations in insolation, can cause the deposits to alternate between net sources and sinks of water vapor. New spacecraft observations of time-dependent temperatures, mineralogy, surface roughness, and volatile stability and migration are helping to document the current reservoirs and exchanges, but more detailed and longer term observations are essential.

6.2. How has the distribution of ice varied in response to astronomically-induced changes in insolation and climate, and over what timescales?

The presence of layering in the PLD suggests that the mass balance of the polar ice (and, thus, the mean annual concentration of atmospheric water vapor) has been strongly modulated by periodic variations in insolation due to changes in the obliquity and orbital elements of Mars (e.g., Laskar et al., 2002; Milkovich and Head, 2005; Levrard et al., 2007). Of these, the martian obliquity exerts the greatest influence, oscillating about its present mean value ($\sim 25°$) with a period of $\sim 1.2 \times 10^5$ years and with an amplitude of oscillation that varies with a period of $\sim 1.3 \times 10^6$ years (Ward, 1992; Laskar and Roubutel, 1993; Touma and Wisdom, 1993; Laskar et al., 2004). Laskar et al. (2004) have produced a precise calculation of the evolution of these parameters over the last 10–20 million years (Laskar et al., 2004); Mars is currently at an anomalously low mean obliquity and amplitude relative to the rest of this time. Although precise predictions for periods prior to $\sim 10^7$ years are not possible due to the mathematically chaotic nature of the solutions, statistical studies of a range of solutions show that obliquity can reach a maximum of 82°, with an average of $\sim 38°$, and with a 63% probability of it being $>60°$ in the past billion years. Large variations in obliquity result in significant changes in regional surface temperature, the stability of polar ice, and the concentration of water vapor in the atmosphere. The exact way in which a driver such as obliquity will affect the climate is difficult to predict. For example, warming might lead to an increase in surface pressure, allowing more dust entrainment and decreasing albedo, hence leading to cooling, or initial cooling may lead to an increased ice stability and a increased albedo, leading to more cooling (Fanale et al., 1982; Jakosky et al., 1993). However, one can make some generalizations. At very low obliquity (~ 10–$15°$), the atmosphere collapses and condenses onto the surface, with distribution controlled by topography (Kreslavsky and Head, 2005). At low obliquity similar to that of today, seasonal temperature fluctuations and mean annual polar temperatures are at a minimum, while equatorial temperatures are at their maximum. Under these conditions, ground ice is stable only at latitudes above ~ 60–$70°$ (Mellon and Jakosky, 1995).

This situation is reversed at times of high obliquity, when long summers of continuous illumination alternate with sunless winters, to produce both extreme seasonal variations and higher mean annual insolation and temperatures at the poles. As the obliquity increases to $\sim 32°$, the higher temperatures at the poles and lower temperatures at the equator result in conditions where ground ice is stable globally within the top few meters of the regolith (Mellon and Jakosky, 1995). However, as the obliquity increases, mean annual temperatures actually rise to the point were ground ice becomes unstable near the poles—a region of instability that expands and includes everywhere but the tropics at an obliquity of $\sim 45°$ (Mischna et al., 2003).

What is the fate of the reservoir of polar volatiles during these periods of higher obliquity? Geological evidence has been presented for the presence of huge tropical mountain glacier deposits on the northwest flanks of Olympus Mons and the Tharsis Montes (Head and Marchant, 2003; Shean et al., 2005; Milkovich et al., 2006; Shean et al., 2007). These deposits have been interpreted to represent the presence of cold-based tropical mountain glaciers in the equatorial regions during the Late Amazonian period of Mars history. Other evidence of non-polar ice-related and glacial deposits is found in the mid-latitude regions in the form of lobate debris aprons (LDA) and lineated valley fill (LVF) (e.g., Squyres, 1978; Squyres, 1979; Lucchitta, 1981; Pierce and Crown, 2003; Head et al., 2005; Head et al., 2006a, 2006b). For the LDA, all agree that ice-assisted processes were fundamental in their formation but disagreement persists on the amount of ice involved, with ideas ranging from ice-assisted creep of talus slopes to debris-covered glaciers (Li et al., 2005). Patterns in the associated LVF clearly implicate glacial processes in the formation and evolution of these systems (e.g., Head et al., 2006a, 2006b). Thus, significant amounts of volatiles have been present in the recent geological past at mid-latitudes as well, and some workers suggest that at least some portion of the ices remain sequestered below an insulating layer of debris.

Evidence also exists for more extensive polar deposits in the past history of Mars. The Hesperian-aged south circumpolar Dorsa Argentea Formation has been interpreted to be an ancient polar deposit underlying the present south PLD, and extending over twice the area of the current PLD (Head and Pratt, 2001). Recent MARSIS data has revealed evidence of layering in this deposit attributed to ice-rich, material (Plaut et al., 2007). These observations not only indicate that there might have been a larger surface volatile reservoir in the past history of Mars, but the MARSIS data suggest that large quantities of volatiles might be sequestered in these types of deposits and removed from the system.

Periglacial features (such as polygons, basketball terrain, boulder fields, scalloped depressions, etc.) also abound in the martian mid-high latitudes (e.g., Mangold, 2005; van Gasselt and Hauber, 2007; Lefort et al., 2007; Mellon et al., 2007). On Earth, patterned ground is a characteristic feature of the periglacial landscape, molded by frost and ice processes, mass movement and wind stripping. Understanding the nature and formation of the patterned ground may be a key to understand the nature of ice/regolith/atmosphere interactions. The inventory of small-scale structures in the martian circumpolar regions is not yet known. It is also crucial to assess the structure of patterned ground (sand- or ice-wedge polygons or sublimation polygons; Mangold, 2005; Marchant and Head, 2007), the possible range of porosities, and whether the current chemistry of the regolith is in equilibrium with modern conditions or represents residual past conditions. New technology and continuing exploration of Mars not only offers the opportunity to map out and date the full range of volatile-related deposits at all latitudes, but also to determine their three-dimensional structure and assess their ice content.

6.3. How are volatiles and dust transported to and from polar regions, and over what time scales?

The continuing development of new and more robust atmospheric general circulation models has permitted simulations to be undertaken which include historical variations in spin-axis and orbital parameters, water vapor and dust content (e.g., Richardson and Wilson, 2002; Haberle et al., 2003; Mischna et al., 2003; Levrard et al., 2007). These models have been useful in beginning to link geological evidence of non-polar volatile deposits with the processes of volatile mobilization, transport, and deposition. For example, Forget et al. (2006) showed that at obliquities of $\sim 45°$, atmospheric water vapor derived from polar deposits would migrate equatorward, rise along the western flanks of Tharsis and encounter the huge Tharsis volcanoes, and undergo upwelling, adiabatic cooling and precipitation of snow, to form the tropical mountain glaciers. Similarly, Madeleine et al. (2007) were able to use GCMs to simulate the northern mid-latitude lineated valley fill glacial deposits by calling on relatively higher amounts of atmospheric dust. These initial results are encouraging in terms of working toward further understanding of broad-scale volatile exchange processes and climate history. Volatile exchange processes at the scale of atmosphere-regolith interactions are crucially important to understanding longer-term exchange processes and clues to these can be obtained using the more recent geological record.

6.4. Are gullies, low-latitude viscous flow features, and mid-to-high-latitude mantling deposits, indicative of changes in climate and volatile exchange?

What happens during periods of increased obliquity, such as from about 0.5 to 2 million years ago, when obliquity often exceeded 30°? Do the ice stability relations simply migrate latitudinally, causing ice to be stable in the regolith at lower latitudes, or is there a more significant change in the distribution of volatiles? MGS spacecraft image and altimetry data have revealed the presence of a very young, latitude-dependent, meters-thick surface layer that subdues underlining topography and has unusual latitude-dependent surface features, such as rounded polygons, gullies, layering, viscous flow features and cryokarst (e.g., Kreslavsky and Head, 2000; Mustard et al., 2001; Milliken et al., 2003). These workers interpreted this deposit, which extends from ∼30° latitude to the poles in both hemispheres, to have been emplaced during recent martian "ice ages" (Head et al., 2003). During these times, Head et al. (2003) argue that water is mobilized from the polar regions and deposited at lower latitudes together with dust to form the mantling layer. Alternating periods of high and low obliquity during the ice age cause sublimation, concentration of dust, decrease in sublimation rates, and the net accumulation of a dusty, mantling, ice-rich deposit. As the amplitude of the obliquity variations decreased in the last several hundred thousand years, the lowest-latitude portions of the deposit (from about 30°–50°) became unstable, undergoing desiccation, and displaying a cryokarst texture due to sublimation and loss of volatiles to polar regions. In this scenario, the viscous flow features and gullies represent special conditions in the stability field linked primarily to crater interiors and resulting variations in slopes and insolation causing viscous flow of accumulated ice (e.g., Milliken et al., 2003) or its local melting (e.g., Costard et al., 2002). This ice-age depositional model is in marked contrast to simple changes in water vapor diffusion and latitudinally migrating ice stability fields predicted to occur during obliquity changes (e.g., Mellon and Jakosky, 1995). The Phoenix mission is scheduled to land at high-latitudes and may be able to distinguish between these two hypotheses by digging into the subsurface and testing for the presence of depositional ice-age layers or pore/secondary ice.

7. Summary

The rich climate and geologic history recorded within the polar deposits has only begun to be discovered, and much remains to be done. The complex relationships between the materials surrounding and underlying the PLD and the PLD themselves must be more completely documented. The ages of the polar materials, a timescale for the layering in both PLD, and the timing of melting and retreat events have yet to be definitively established. The mass balances and current absolute accumulation rates of the residual caps and PLD have never been directly measured and must, for now, be inferred by various means. Correlations between the stratigraphic records at both poles and the relationship between those records and climate change are yet in the preliminary stages. And the catalog of lower latitude ice-related deposits, possibly fed by waning of the polar deposits, is still growing. Ongoing observations by Mars Odyssey, Mars Reconnaissance Orbiter, and Mars Express and future exploration by Phoenix will provide crucial further insight to the major questions in Mars polar science outlined here. But landed missions dedicated to the study of the residual caps and PLD are likely necessary to provide more definitive answers.

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References

Aharonson, O., Zuber, M., Smith, D., Neumann, G., Feldman, W., Prettyman, T., 2004. Depth, distribution, and density of CO_2 deposition on Mars. J. Geophys. Res. 109, doi:10.1029/2003JE002223.

Anguita, F., Babín, R., Benito, G., Gomez, D., Collado, A., Rice, J., 2000. Chasma Australe, Mars: Structural framework for a catastrophic outflow origin. Icarus 144, 302–312.

Barker, E., Schorn, R., Woszczyk, A., Tull, R., Little, S., 1970. Mars: Detection of atmospheric water vapor during the southern hemisphere spring and summer season. Science 170, 1308–1310.

Bass, D., Herkenhoff, K., Paige, D., 2000. Variability of Mars' north polar water ice cap. I. Analysis of Mariner 9 and Viking Orbiter imaging data. Icarus 144, 382–396.

Benito, G., Mediavilla, F., Fernandez, M., Marquez, A., Martinez, J., Anguita, F., 1997. Chasma Boreale, Mars: A sapping and outflow channel with a tectono-thermal origin. Icarus 129, 528–538.

Bibring, J.-P., and 12 colleagues, 2004. Perennial water ice identified in the south polar cap of Mars. Nature 428, 627–630.

Bibring, J.-P., and 10 colleagues, 2005. Mars surface diversity as revealed by the OMEGA/Mars Express observations. Science 307, 1576–1581.

Boynton, W., and 24 colleagues, 2002. Distribution of hydrogen in the near surface of Mars: Evidence for subsurface ice deposits. Science 297 (5578), 81–85.

Byrne, S., Ingersoll, A., 2003. A sublimation model for martian south polar ice features. Science 299, 1051–1053.

Byrne, S., Ivanov, A., 2004. Internal structure of the martian south polar layered deposits. J. Geophys. Res. 108 (E11), doi:10.1029/2004JE002267.

Byrne, S., Murray, B., 2002. North polar stratigraphy and the paleo-erg of Mars. J. Geophys. Res. 107 (E6), doi:10.1029/2001JE001615.

Byrne, S., Zuber, M., 2006. Landscape evolution and the reincarnation of the southern residual ice cap. Paper presented at Fourth International Mars Polar Science and Exploration Conference. Abstract 8035.

Byrne, S., Herkenhoff, K., Russell, P., Hansen, C., McEwen, A., HiRISE team, 2007. Preliminary HiRISE polar geology results. Lunar Planet. Sci. 38. Abstract 1930.

Calvin, W., Pocock, J., James, P., Lee, S, Thomas, P., Murchie, S. Seelos, F., Langevin, Y., Titus, T., McGuire, P., 2007. The nature and mobility of bright patches in the north residual ice cap from MARCI, CTX. OMEGA, and CRISM. Paper presented at 7th International Conference on Mars. Abstract 3267.

Cantor, B., Malin, M., Edgett, K., 2002. Multiyear Mars Orbiter Camera (MOC) observations of repeated martian weather phenomena during the northern summer season. J. Geophys. Res. 107, 3-1–3-8.

Christensen, P.R., and 21 colleagues, 2003. Morphology and composition of the surface of Mars: Mars Odyssey THEMIS results. Science 300 (5628), 2056–2061.

Clifford, S., 1987. Polar basal melting on Mars. J. Geophys. Res. 92, 9135–9152.

Clifford, S., 1993. A model for the hydrologic and climatic behavior of water on Mars. J. Geophys. Res. 98, 10973–11016.

Clifford, S., and 52 colleagues, 2000a. The state and future of Mars polar science and exploration. Icarus 144, 210–242.

Clifford, S.M., Fisher, D.A., Rice, J.W., 2000b. Introduction to the Mars Polar Science Special Issue: Exploration platforms, technologies, and potential future missions. Icarus 144 (2), 205–209.

Clifford, S.M., Thorsteinsson, T., Bjornsson, H., Fisher, D.A., Paige, D.A., 2001. Introduction to the Second Mars Polar Science Special Issue. Icarus 154 (1), 1–2.

Clifford, S.M., Doran, P.T., Fisher, D.A., Herd, C.D.K., 2005. Third Mars polar science special issue: Key questions, needed observations, and recommended investigations. Icarus 174 (2), 291–293.

Colaprete, A., Haberle, R., Hollingsworth, J., Kieffer, H., Titus, T., 2005. Albedo of the south pole on Mars determined by topographic forcing of atmospheric dynamics. Nature 435, 184–187.

Costard, F., Forget, F., Mangold, N., Peulvast, J.–P., 2002. Formation of recent martian debris flows by melting of near-surface ground ice at high obliquity. Science 295 (5552), 110–113.

Cutts, J., Lewis, B., 1982. Models of climate cycles recorded in martian polar layered deposits. Icarus 50, 216–244.

Cutts, J., Blasius, K., Briggs, G., Carr, M., Greeley, R., Masursky, H., 1976. North polar region of Mars: Imaging results from Viking 2. Science 194, 1329–1337.

Douté, S., Schmitt, S., Langevin, Y., Bibring, J.-P., Altieri, F., Bellucci, G., Gondet, B., Poulet, F., MEx OMEGA team, 2006. South Pole of Mars: Nature and composition of the icy terrains from Mars Express OMEGA observations. Planet. Space Sci. 55 (1-2), 113–133.

Edgett, K., Williams, R., Malin, M., Cantor, B., Thomas, P., 2003. Mars landscape evolution: Influence of stratigraphy on geomorphology in the north polar region. Geomorphology 52, 289–297.

Fanale, F., 1976. Martian volatiles: Their degassing history and geochemical fate. Icarus 28, 179–202.

Fanale, F., Salvail, J., Banerdt, W., Saunders, R.S., 1982. Mars: The regolith–atmosphere–cap system and climate change. Icarus 50, 381–407.

Fanale, F., Salvail, J., Zent, A., Postawko, S., 1986. Global distribution and migration of subsurface ice on Mars. Icarus 67, 1–18.

Farmer, C., Doms, P., 1979. Global and seasonal variations of water vapor on Mars and the implications for permafrost. J. Geophys. Res. 84, 2881–2888.

Farmer, C., Davies, D., LaPorte, D., 1976. Mars: Northern summer ice cap—Water vapor observations from Viking 2. Science 194, 1339–1340.

Feldman, W., and 12 colleagues, 2002. Global distribution of Neutrons from Mars: Results from Mars Odyssey. Science 297, 75–78.

Feldman, W., Prettyman, T.H., Boynton, W.V., Murphy, J.R., Squyres, S., Karunatillake, S., Maurice, S., Tokar, R.L., McKinney, G.W., Hamara, D.K., 2003. CO_2 frost cap thickness on Mars during northern winter and spring. J. Geophys. Res. 108 (E9), doi:10.1029/2003JE002101.

Fenton, L., Herkenhoff, K., 2000. Topography and stratigraphy of the martian north polar layered deposits using photoclinometry, stereogrammetry, and MOLA altimetry. Icarus 147, 433–443.

Fishbaugh, K., Head, J., 2000. North polar region of Mars: Topography of circumpolar deposits from Mars Orbiter Laser Altimeter (MOLA) data and evidence for asymmetric retreat of the polar cap. J. Geophys. Res. 105, 22455–22486.

Fishbaugh, K., Head, J., 2002. Chasma Boreale, Mars: Topographic characterization from Mars Orbiter Laser Altimeter data and implications for mechanisms of formation. J. Geophys. Res. 107 (E3), doi:10.1029/2000JE001351.

Fishbaugh, K., Head, J., 2005. Origin and characteristics of the Mars north polar basal unit and implications for polar geologic history. Icarus 174 (2), 444–474.

Fishbaugh, K., Hvidberg, C., 2006. Martian north polar layered deposits stratigraphy: Implications for accumulation rates and flow. J. Geophys. Res. 111, doi:10.1029/2005JE002571. E06012.

Fishbaugh, K., Herkenhoff, K., Byrne, S., Russell, P., McEwen, A., Hansen, C., HiRISE team, 2007a. HiRISE observations of north polar stratigraphy and implications for geologic history. Paper presented at 7th International Conference on Mars. Abstract 3194.

Fishbaugh, K., Poulet, F., Langevin, Y., Chevrier V., Bibring, J.-P., 2007b. On the origin of gypsum in the Mars North Polar region. J. Geophys. Res. 112 doi:10.1029/2006JE002862. E06012.

Fisher, D., 2000. Internal layers in an "accublation" ice cap: A test for flow. Icarus 144, 289–294.

Fisher, D., 2007. Mars' water isotope (D/H) history in the strata of the North Polar Cap: Inferences about the water cycle. Icarus 187, 430–441.

Forget, F., Pierrehumbert, R., 1997. Warming early Mars with carbon dioxide clouds that scatter infrared radiation. Science 278 (5341), 1273–1276.

Forget, F., Hourdin, F., Talagrand, O., 1998. CO_2 snowfall on Mars: Simulation with a general circulation model. Icarus 131, 302–316.

Forget, F., Haberle, R., Montmessin, F., Levrard, B., Head, J., 2006. Formation of glaciers on Mars by atmospheric precipitation at high obliquity. Science 311 (5759), 368–371.

Garvin, J., Sakimoto, S., Frawley, J., Schnetzler, C., Wright, H., 2000. Topographic evidence for geologically recent near-polar volcanism on Mars. Icarus 145 (2), 648–652.

Greve, R., Klemann, V., Wolf, D., 2003. Ice flow and isostasy of the north polar cap of Mars. Planet. Space Sci. 51, 193–204.

Haberle, R., Murphy, J., Schaeffer, J., 2003. Orbital change experiments with a Mars General Circulation Model. Icarus 161, 66–89.

Hale, A., Bass, D., Tamppari, L., 2005. Monitoring the perennial martian northern polar cap with MGS MOC. Icarus 174, 502–512.

Hansen, G., and 12 colleagues, 2005. PFS-MEX observation of ices in the residual south polar cap of Mars. Planet. Space Sci. 53, 1089–1095.

Head, J., Marchant, D., 2003. Cold-based mountain glaciers on Mars: Western Arsia Mons. Geology 31 (7), 641–644.

Head, J., Pratt, S., 2001. Extensive Hesperian-aged south polar ice sheet on Mars: Evidence for massive melting and retreat, and lateral flow and ponding of meltwater. J. Geophys. Res. 106 (E6), 12275–12300.

Head, J., Mustard, J., Kreslavsky, M., Milliken, R., Marchant, D., 2003. Recent ice ages on Mars. Nature 426, 797–802.

Head, J.W., Neukum, G., Jaumann, R., Hiesinger, H., Hauber, E., Carr, M., Masson, P., Foing, B., Hoffmann, H., Kreslavsky, M., 2005. Tropical to mid-latitude snow and ice accumulation, flow and glaciation on Mars. Nature 434, 346–351.

Head, J., Nahm, A., Marchant, D., Neukum, G., 2006b. Modification of the dichotomy boundary on Mars by Amazonian mid-latitude regional glaciation. Geophys. Res. Lett. 33 (8), doi:10.1029/2005GL024360.

Head, J., Marchant, D., Agnew, M., Fassett, C., Kreslavsky, M., 2006a. Extensive valley glacier deposits in the northern mid-latitudes of Mars: Evidence for Late Amazonian obliquity-driven climate change. Earth Planet. Sci. Lett. 241 (3–4), 663–671.

Herkenhoff, K., Murray, B., 1990. Color and albedo of the south polar layered deposits on Mars. J. Geophys. Res. 95, 14511–14529.

Herkenhoff, K., Plaut, J., 1999. Ages and resurfacing rates of the polar layered deposits. Paper presented at The Fifth International Conference on Mars 5. Abstract 6065.

Herkenhoff, K., Soderblom, L., Kirk, R., 2003. Stratigraphy and structure of the south polar layered deposits on Mars. Paper presented at 3rd Mars Polar Science and Exploration Conference. Abstract 8030.

Herkenhoff, K., Byrne, S., Russell, P., Fishbaugh, K., McEwen, A., 2007. HiRISE observations of the north polar region of Mars. Science 317, 1711–1715.

Herschel, W., 1784. On the remarkable appearances of the polar regions of the planet Mars, the inclination of its axis, the position of its poles, and its spheroidical figure; with a few hints relating to its real diameter and atmosphere. Philos. Trans. 24, 233–273.

Hodges, C., Moore, H., 1994, Atlas of volcanic landforms on Mars. in: United States Geological Survey Professional Paper 1534.

Houben, H., Haberle, R., Young, R., Zent, A., 1997. Modeling the martian seasonal water cycle. J. Geophys. Res. 102, 9069–9083.

Howard, A., 2000. The role of eolian processes in forming surface features of the martian polar layered deposits. Icarus 144, 267–288.

Howard, A., Cutts, J., Blasius, K., 1982. Stratigraphic relationships within martian polar cap deposits. Icarus 50, 161–215.

Hvidberg, C., 2003. Relationship between topography and flow in the north polar cap on Mars. Ann. Glaciol. 37, 363–369.

Jakosky, B., Henderson, B., Mellon, M., 1993. The Mars water cycle at other epochs: Recent history of the polar caps and layered terrain. Icarus 102, 286–297.

Jakosky, B., Zent, A., Zurek, R., 1997. The Mars water cycle: Determining the role of exchange with the regolith. Icarus 1, 87–95.

Jakosky, B., Mellon, M., Varnes, S., Feldman, W., Boynton, W., Haberle, R., 2005. Mars low-latitude neutron distribution: Possible remnant near-surface water ice and a mechanism for its recent emplacement. Icarus 175 (1), 58–67.

James, P.B., Cantor, B.A., 2001. Martian North Polar Cap recession: 2000 Mars Orbiter Camera observations. Icarus 154 (1), 131–144.

Kieffer, H., 1979. Mars south polar spring and summer temperatures: A residual CO_2 frost. J. Geophys. Res. 84, 8263–8288.

Kieffer, H., 1990. H_2O grain size and the amount of dust in the Mars' residual north polar cap. J. Geophys. Res. 95, 1481–1493.

Kieffer, H., Titus, T., 2001. TES mapping of Mars' north seasonal cap. Icarus 154, 162–180.

Kieffer, H., Chase, S., Miner Jr., T.M.E., Palluconi, F., 1976. Martian north pole summer temperatures: Dirty water ice. Science 194, 1341–1344.

Kolb, E., Tanaka, K., 2001. Geologic history of the polar regions of Mars based on Mars Global Surveyor data. II. Amazonian period. Icarus 154, 22–39.

Kolb, E., Tanaka, K., 2006. Accumulation and erosion of south polar layered deposits in the Promethei Lingula region, Planum Australe, Mars. Mars 2.

Kolb, E., Tanaka, K., 2007. Accumulatioin and erosion of south polar layered deposits in the Australe Scopuli region, Planum Australe, Mars. Icarus, submitted for publication.

Koutnik, M., Byrne, S., Murray, B., 2002. South polar layered deposits of Mars: The cratering record. J. Geophys. Res. 107 (E11), doi:10.1029/2001JE001805.

Kreslavsky, M., Head, J., 2000. Kilometer-scale roughness of Mars: Results from MOLA data analysis. J. Geophys. Res. 105, 26695–26711.

Kreslavsky, M., Head, J., 2005. Mars at very low obliquity: Atmospheric collapse and the fate of volatiles. Geophys. Res. Lett. 32, doi:10.1029/2005GL022645. L12202.

Langevin, Y., Poulet, F., Bibring, J.-P., Schmitt, B., Douté, S., Gondet, B., 2005. Summer evolution of the north polar cap of Mars as observed by OMEGA/Mars Express. Science 307 (5715), 1581–1584.

Laskar, J., Roubutel, P., 1993. The chaotic obliquity of the planets. Nature 361, 608–612.

Laskar, J., Levrard, B., Mustard, F., 2002. Orbital forcing of the martian polar layered deposits. Nature 419, 375–377.

Laskar, J., Correia, A., Gastineau, M., Joutel, F., Levard, B., Robutel, P., 2004. Long term evolution and chaotic diffusion of the insolation quantities on Mars. Icarus 170, 343–364.

Lefort, A., Russell, P., Thomas, N., HiRISE team, 2007. Periglacial features in the martian mid-latitudes, insight from HiRISE. Paper presented at 2nd International Workshop: Exploring Mars and its Earth Analogs. Abstract 89.

Leighton, R., Murray, B., 1966. Behavior of carbon dioxide and other volatiles on Mars. Science 153, 136–144.

Levrard, B., Forget, F., Montmessin, F., Laskar, J., 2007. Recent formation and evolution of northern martian polar layered deposits as inferred from a Global Climate Model. J. Geophys. Res. 112, doi:10.1029/2006/JE002772. E06012.

Li, H., Robinson, M., Jurdy, D., 2005. Origin of martian northern hemisphere mid-latitude lobate debris aprons. Icarus 176 (2), 382–394.

Lucchitta, B., 1981. Mars and Earth—Comparison of cold-climate features. Icarus 45, 264–303.

Madeleine, J.B., Forget, F., Head, J.W., Levrard, B., Montmessin, F., 2007. Mars: A proposed climatic scenario for northern mid-latitude glaciation. Lunar Planet. Sci. 38. Abstract 1338.

Malin, M., Edgett, K., 2001. Mars Global Surveyor Mars Orbiter Camera: Interplanetary cruise though primary mission. J. Geophys. Res. 106 (E10), 23429–23570.

Mangold, N., 2005. High latitude patterned ground on Mars: Classification, distribution, and climatic control. Icarus 174, 336–359.

Marchant, D., Head, J., 2007. Antarctic Dry Valleys: Microclimate zonation, variable geomorphic processes, and implications for assessing climate change on Mars. Icarus 192, 187–222.

Melchiorri, R., and 10 colleagues, 2006. Water vapor mapping on Mars using OMEGA/Mars Express. Planet. Space Sci. 55 (3), 333–342.

Mellon, M., Jakosky, B., 1993. Geographic variations in the thermal and diffusive stability of ground ice on Mars. J. Geophys. Res. 98, 3345–3364.

Mellon, M., Jakosky, B., 1995. The distribution and behavior of martian ground ice during past and present epochs. J. Geophys. Res. 100, 11781–11799.

Mellon, M., Feldman, W., Prettyman, T., 2004. The presence and stability of ground ice in the southern hemisphere of Mars. Icarus 169 (2), 324–340.

Mellon, M., Searls, M., Martinez-Alonso, S., HiRISE team, 2007. HiRISE observations of patterned ground on Mars. Paper presented at 7th International Conference on Mars. Abstract 3285.

Milkovich, S., Head, J., 2005. North polar cap of Mars: Polar layered deposit characterization and identification of a fundamental climate signal. J. Geophys. Res. 110 (E05), doi:10.1029/2004JE002349.

Milkovich, S., Head, J., 2006. Surface textures of Mars' north polar layered deposits: A framework for interpretation and future exploration. Mars 2, 21–45.

Milkovich, S., Plaut, J., 2007. Stratigraphic analysis of the south polar layered deposits using radar and imagery. Paper presented at Seventh International Conference on Mars. Abstract 3197.

Milkovich, S., Head, J., Marchant, D., 2006. Debris-covered piedmont glaciers along the northwest flank of the Olympus Mons scarp: Evidence for low-latitude ice accumulation during the Late Amazonian of Mars. Icarus 181 (2), 388–407.

Milliken, R., Mustard, J., Goldsby, D., 2003. Viscous flow features on the surface of Mars: Observations from high-resolution Mars Orbiter Camera (MOC) images. J. Geophys. Res. 108 (E6), doi:10.1029/2003JE002051.

Mischna, M., Richardson, M., Wilson, R., McCleese, D., 2003. On the orbital forcing of martian water and CO_2 cycles: A general circulation model study with simplified volatile schemes. J. Geophys. Res. 108 E(06), doi:10.1029/2003JE002051.

Murchie, S., CRISM team, 2007. First results from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM). Lunar Planet. Sci. 38. Abstract 1472.

Murray, B., Koutnik, M., Byrne, S., Soderblom, L., Herkenhoff, K., Tanaka, K., 2001. Preliminary geological assessment of the northern edge of Ultimi Lobe, Mars South Polar Layered Deposits. Icarus 154, 80–97.

Mustard, J., Cooper, C., Rifkin, M., 2001. Evidence for recent climate change on Mars from the identification of youthful near-surface ground ice. Nature 412 (6845), 411–414.

Novak, R., Mumma, M., Lee, S., Ivanov, L., Bonev, B., Villanueva, G., 2005. Mapping of D/H and ozone in the martian atmosphere near perihelion. Bull. Am. Astron. Soc. 35 (3), 660.

Pathare, A., Murray, B., 2006. Five questionable assertions about the recent mass balance of the martian North Polar Layered Deposits. Paper presented at 4th International Mars Polar Science and Exploration Conference. Abstract 8087.

Pathare, A., Paige, D., 2005. The effects of martian orbital variations upon the sublimation and relaxation of north polar troughs and scarps. Icarus 174, 419–443.

Pathare, A., Koutnik, M., Murray, B., Marshall, S., 2006. Glacial flow modeling of the martian north polar layered deposits. Paper presented at 4th Mars Polar Science and Exploration Conference. Abstract 8072.

Phillips, R., and 26 colleagues, 2008. Mars North Polar Deposits: Stratigraphy, age, and geodynamical response. Science 320 (5880), 1182–1185. 10.1126/science.1157546.

Picardi, G., and 33 colleagues, 2005. Radar soundings of the subsurface of Mars. Science 310 (5756), 1925–1928.

Pierce, T., Crown, D., 2003. Morphologic and topographic analyses of debris aprons in the eastern Hellas region, Mars. Icarus 163 (1), 46–65.

Piquex, S., Byrne, S., Richardson, M., 2003. Sublimations of Mars' southern seasonal CO_2 ice cap and the formation of spiders. J. Geophys. Res. 108 (E8), doi:10.1029/2002JE002007.

Plaut, J., Kahn, R., Guinness, E., Arvidson, R., 1988. Accumulation of sedimentary debris in the south polar region of Mars and implications for climate history. Icarus 75, 357–377.

Plaut, J., Picardi, G., Safaeinili, A., Ivanov, A.B., Milkovich, S.M., Cicchetti, A., Kofman, W., Mouginot, J., Farrell, W.M., Phillips, R.J., 2007. Subsurface radar sounding of the south polar layered deposits of Mars. Sci. Express, doi:10.1126/science.1139672.

Richardson, M., Wilson, R., 2002. Investigation of the nature and stability of the martian seasonal water cycle with a general circulation model. J. Geophys. Res. 107, doi:10.1029/2001JE001536.

Russell, P., Byrne, S., Herkenhoff, K., Fishbaugh, K., McEwen, A., HiRISE team, 2007. North polar basal stratigraphy and active mass-wasting processes with implications for polar scarp evolution. Paper presented at 7th International Mars Conference. Abstract 3377.

Schorghofer, N., Aharonson, O., 2005. Stability and exchange of subsurface ice on Mars. J. Geophys. Res. 110, doi:10.1029/2004JE002350. (E05003).

Shean, D., Head, J., Marchant, D., 2005. Origin and evolution of a cold-based tropical mountain glacier on Mars: The Pavonis Mons fan-shaped deposit. J. Geophys. Res. 110, doi:10.1029/2004JE002360.

Shean, D., Head, J., Fastook, J., Marchant, D., 2007. Recent glaciation at high elevations on Arsia Mons, Mars: Implications for the formation and evolution of large tropical mountain glaciers. J. Geophys. Res. 112, doi:10.1029/2006JE002761. E03004.

Sizemore, H., Mellon, M., 2006. Effects of subsurface heterogeneity on martian ground-ice stability and orbital estimates of ice table depth. Icarus 185 (2), 358–369.

Smith, D.E., Zuber, M.T., Neumann, G.A., 2001. Seasonal variations of snow depth on Mars. Science 294 (5549), 2141–2146.

Smoluchowski, R., 1968. Retention of ice. Science 159, 1350–1384.

Squyres, S., 1978. Martian fretted terrain—Flow of erosional debris. Icarus 34, 600–613.

Squyres, S., 1979. The distribution of lobate debris aprons and similar flows on Mars. J. Geophys. Res. 84, 8087–8096.

Squyres, S.W., and 50 colleagues, 2004a. The Opportunity Rover's Athena science investigation at Meridiani Planum. Mars. Science 306, 1698–1703.

Squyres, S.W., and 18 colleagues, 2004b. In situ evidence for an ancient aqueous environment at Meridiani Planum, Mars. Science 306, 1709–1714.

Tanaka, K., 2005. Geology and insolation-driven climatic history of Amazonian north polar materials on Mars. Nature 437, 991–994.

Tanaka, K., Bourke, M., 2007. Stratigraphy and landscape evolution of Chasma Boreale, Mars. Lunar Planet. Sci. 38. Abstract 1856.

Tanaka, K., Kolb, E., 2001. Geologic history of the polar regions of Mars based on Mars Global Surveyor data. I. Noachian and Hesperian Periods. Icarus 154, 3–21.

Tanaka, K.L., Rodriguez, J.A.P., Skinner Jr., J.A., Bourke, M.C., Fortezzo, C.M., Herkenhoff, K.E., Kolb, E.J., Okubo, C.H., 2008. North polar region of Mars: Advances in stratigraphy, structure, and erosional modification. Icarus 196, 318–358.

Thomas, P., Weitz, C., 1989. Sand dune materials and polar layered deposits on Mars. Icarus 81, 185–215.

Tillman, J., Johnson, N., Gettorp, P., Percival, D., 1993. The martian annual atmospheric pressure cycle: Years without great dust storms. J. Geophys. Res. 98, 10963–10971.

Touma, J., Wisdom, J., 1993. The chaotic obliquity of Mars. Science 259 (5099), 1294–1297.

van Gasselt, S., Hauber, E., 2007. Cold-climate landforms and processes on Mars. Paper presented at European Geophysical Union General Assembly. Abstract EGU2007-A-11532.

Ward, W., 1992. Long-term orbital and spin dynamics of Mars. In: Kieffer, H., Jakosky, B., Snyder, C., Matthews, M. (Eds.), Mars. University of Arizona Press, Tucson, AZ, pp. 298–320.

Warren, S., Wiscombe, W., 1980. A model for the spectral albedo of snow. II. Snow containing atmospheric aerosols. J. Atmos. Sci. 37, 2734–2745.

Winebrenner, D., Koutnik, M., Waddington, E., Pathare, A., Murray, B., Byrne, S., 2006. Shaping the NPLD: Evidence for recent change in dominant process. Paper presented at 4th Mars Polar Science and Exploration Conference. Abstract 8055.

Wrobel, K.E., Schultz, P.H., 2004. Effect of planetary rotation on distal tektite deposition on Mars. J. Geophys. Res. 109, doi:10.1029/2004JE002250. E05005.

Zent, A., Fanale, F., Salvail, J., Postawko, S., 1986. Distribution and state of H_2O in the high-latitude shallow subsurface of Mars. Icarus 67, 19–36.

Zuber, M.T., and 20 colleagues, 1998. Observations of the North Polar Region of Mars from the Mars Orbiter Laser Altimeter. Science 282 (5396), 2053–2060.