## CONTENTS — H through K

- **Evaluation the Phoenix Region B Landing Site Rock Coverage from Available Radar Data**  
  *A. F. C. Haldemann and B. J. Butler* ................................................................. 8057

- **Regional Polar Glaciation in the Hesperian Period of the History of Mars: The South Circumpolar Dorsa Argentea Formation as an Ancient Ice Sheet Remnant**  
  *J. W. Head and D. R. Merchant* ......................................................................... 8103

- **Chronos: A Journey Through Martian History**  
  *M. H. Hecht and Chronos Team* ......................................................................... 8096

- **Understanding Ambiguities in Backscattered Orbital Radar Sounding Data from the Martian Polar Layered Deposits Using Finite Difference Time Domain Simulations**  
  *E. Heggy, S. M. Clifford, and D. Card* ................................................................. 8107

  *E. Heggy, S. M. Clifford, A. Younsi, J. L. Miane, R. Carley, and R. V. Morris* .......... 8105

- **Imaging the Polar Regions of Mars with HiRISE**  
  *K. Herkenhoff, S. Byrne, C. Hansen, V. Gulick, and A. McEwen* ......................... 8009

- **Mars Polar Geologic Nomenclature: What are the Caps?**  
  *K. E. Herkenhoff, S. Byrne, and K. L. Tanaka* ..................................................... 8034

- **Stratigraphy and Structure of the North Polar Layered Deposits on Mars**  
  *K. E. Herkenhoff, L. A. Soderblom, R. L. Kirk, L. Keszthelyi, T. Becker, and L. Weller* 8027

- **Arctic Analogue Science as Part of an Integrated Canadian Strategy for Mars Exploration**  
  *V. J. Hipkin, A. Berinstain, D. Laurin, A. Ouellet, M. Lebeuf, G. R. Osinski, and R. Léveillé* 8078

- **Airborne Radar Sounding in Antarctica: Analogs for SHARAD Data over the Polar Layered Terrains and Valley Glaciers of Mars**  

- **The North Polar Layered Deposits, Mars: Topography, Flow and Implications for Timescales**  
  *C. S. Hvidberg* ...................................................................................................... 8101

- **The North Polar Layered Deposits, Mars: Sublimation Rates and Recent Evolution**  
  *C. S. Hvidberg, N. B. Karlsson, and D. Tyler Jr.* .................................................... 8075

- **Observations of the Layering Structure in the South Polar Layered Deposits with the MARSIS Instrument**  
  *A. B. Ivanov, A. Safaeinili, J. J. Plaut, S. M. Milkovich, and G. Picardi* ................. 8070

- **Interannual Variability in Residual South Cap Albedo: MOC Observations**  
  *P. B. James, B. P. Bonev, and M. J. Wolff* ............................................................ 8020

- **Displacement of Martian Surface Due to Seasonal Surface Mass Redistribution and Its Detection from Lander-Orbiter-Earth Radio Links**  
  *Ö. Karatekin, J. Hagedoorn, T. Van Hoolst, and V. Dehant* .................................... 8039
Mass and Density of Seasonal Polar Deposits from Time-Variable Gravity Data
Ö. Karatekin, T. Van Hoolst, and V. Dehant .................................................................................. 8067

Ice Caps on Mars and on Kilimanjaro: Potential for Analogies
G. Kaser, N. Cullen, and T. Mölg ............................................................................................... 8099

The Robotic Arm Camera (RAC)
H. U. Keller, W. J. Markiewicz, S. Hviid, M. T. Lemmon, P. H. Smith, 
R. V. Morris, and Phoenix Science Team .................................................................................. 8042

Geologic Mapping Results of the Australe Scopuli Region Within Planum Australe, Mars 
E. J. Kolb and K. L. Tanaka ..................................................................................................... 8085

Conditions for Water Ice Accumulation Within Craters on Mars
K. J. Kossacki, W. J. Markiewicz, G. Poutyankina, G. Neukum, and HRSC Team .................. 8016

The 2007 Phoenix Mars Scout Wet Chemistry Laboratory: Studying the Chemistry of the Polar Soil/Ice
S. P. Kounaves, M. H. Hecht, and Phoenix Team ...................................................................... 8068

Inferring Spatial Patterns of Accumulation from Radar Internal Layers
M. R. Koutnik, E. D. Waddington, and D. P. Winebrenner ...................................................... 8036

Inner Structure of the Polar Layered Deposits on Mars: Relation to the Climate Record
M. A. Kreslavsky and J. W. Head ............................................................................................... 8028
EVALUATING THE PHOENIX REGION B LANDING SITE ROCK COVERAGE FROM AVAILABLE RADAR DATA. A. F. C. Haldemann, and B. J. Butler, Jet Propulsion Laboratory, California Institute of Technology, (albert.haldemann@jpl.nasa.gov), 4800 Oak Grove Dr., M/S 183-301, Pasadena, CA 91109, National Radio Astronomy Observatory, P.O. Box O, Socorro, NM 87801.

Introduction: The Phoenix Mars Scout mission, to be launched in August 2007 down-selected their choice of landing target area to “Region B” in December 2005 [1]. Region B, bordered by latitudes 65°N and 72°N and longitudes 120°E and 140°E, was selected for landing safety in addition to the science potential of landing sites within it.

A critical consideration for landing safety is rock coverage: the rock abundance must be < 18%. For Phoenix a difficulty arises in the determination of rock abundance at high latitudes. The dataset that has been used for rock coverage evaluation of previous, near-equatorial, landing site selections is the 1° resolution rock abundance derived IRTM thermal-inertias [2]. At high-latitudes the technique of deriving rock abundance from thermal inertia becomes more uncertain because there is less diurnal thermal contrast. Nevertheless, the rock abundance estimate from this technique for Region B is 18%, like the Viking Lander 2 (VL2) site, which also serves as a geomorphic analog for Region.

Other techniques have been applied to measuring the Phoenix Region B rock abundance. Rock abundance and boulder characteristics have been determined using MOC data [3-4]. Further evaluations have been attempted with bi-static radar experiments [5-6]. In [5] UHF transmissions from Earth were received by the Mars Odyssey spacecraft. The direct signal is compared to the echo specularly reflected off the Martian surface to evaluate dielectric constant and possible near-surface ice layers. In theory a dielectric constant measurement can distinguish between loose soil ($\varepsilon \approx 2$) and dense rock ($\varepsilon \geq 5$). Unfortunately the polarization performance of the UHF transmitter limited the interpretability of the data (J. Callas, personal communication). The bi-static radar experiments carried out with the Mars Express (MEX) spacecraft [6] worked better, and sampled the VL2 region bistatically at both 3.5 cm and 12.6 cm wavelength scales. However, MEX will only sample Region B directly in June 2006. The VL2 MEX data, while generally in family with previous observations, were puzzling for their wavelength dependence, and warrant further review. Notably, the observation date of the data presented in [6] was Dec. 12, 2005, when Ls~330°, in mid-to late northern winter, so surface frost deposits may be part of the explanation.

Goldstone-VLA Radar Data: We report on the analysis of existing Earth-based radar data from the Goldstone-Very Large Array (GVLA) combination. Unlike the radar evaluations of near-nadir quasi-specular backscatter for the Viking, Pathfinder (MPF), and Mars Exploration Rovers (MER), the backscatter radar data available for Phoenix only exist at high incidence angles. GVLA data are generated by continuous wave transmission from the Goldstone Solar System Radar (GSSR), reception at the VLA, and interferometric imaging of the received echoes. GVLA experiments generate a global dataset at about 60 km resolution of diffuse (incidence angle, $\theta$>30°) radar scattering behaviors [7]. The GVLA diffuse scattering behaviors are reported as parameters, $A_P$ and $n_P$, of cosine scattering models which describe the reflectivity, $\sigma^0_{\text{P}}$, as a function of $\theta$:

$$\sigma^0_{\text{P}}(\theta) = A_P \cos^{n_P} \theta$$

(1)

where the subscript P denotes either the same-sense, or opposite sense of circular polarization ($P=\text{SC}$ and $P=\text{OC}$ respectively). In radar jargon, these polarization channels are also referred to as ‘depolarized’ and ‘polarized’ respectively. The total cross-sections are

$$C_P = A_P / (n_P+2)$$

(2)

The GVLA parameters for proposed and past landing sites are listed in Table 1. The uncertainty on the high-latitude fits is larger than for previous landing sites because the fits are based only on high incidence angle data, as shown in Figure 1.

![Figure 1. GVLA SC data at Region B, with two equal fits (green). The SC reflectivity is higher than at VL2 (blue).](image-url)
Extracting Rock Coverage: Baron et al. [8] developed a technique that ties the diffuse backscatter to the rock coverage. The method calculates backscatter from each individual rock in a distribution and obtains the full surface cross section, by adding up the backscatter of all the rocks. This plausibly suggests linearity of surface rock abundance influence on reflectivity. Among the fit parameters to eq.(1), and the cross sections of eq.(2), only the OC cross-section, $C_{OC}$ shows correlation to the IRTM rock coverage (also Table 1), as shown in Figure 2.

The fits were done precisely at the specified center location with a spatial resolution of ~60km.

<table>
<thead>
<tr>
<th>Landing Region</th>
<th>Latitude</th>
<th>Longitude</th>
<th>IRTM Rock %</th>
<th>$A_{SC}$</th>
<th>$n_{SC}$</th>
<th>$C_{SC}$</th>
<th>$A_{OC}$</th>
<th>$n_{OC}$</th>
<th>$C_{OC}$</th>
<th>$\mu$</th>
</tr>
</thead>
<tbody>
<tr>
<td>VL1</td>
<td>22.3°N</td>
<td>48.0°W</td>
<td>16</td>
<td>0.062</td>
<td>0.76</td>
<td>0.022</td>
<td>0.078</td>
<td>0.12</td>
<td>0.037</td>
<td>0.60</td>
</tr>
<tr>
<td>VL2</td>
<td>47.7°N</td>
<td>225.7°W</td>
<td>17</td>
<td>0.041</td>
<td>0.03</td>
<td>0.020</td>
<td>0.161</td>
<td>0.14</td>
<td>0.075</td>
<td>0.27</td>
</tr>
<tr>
<td>MPF</td>
<td>19.3°N</td>
<td>33.6°W</td>
<td>18</td>
<td>0.038</td>
<td>0.20</td>
<td>0.017</td>
<td>0.085</td>
<td>0.17</td>
<td>0.039</td>
<td>0.43</td>
</tr>
<tr>
<td>Gusev</td>
<td>14.8°S</td>
<td>184.9°W</td>
<td>8</td>
<td>0.082</td>
<td>0.78</td>
<td>0.029</td>
<td>0.149</td>
<td>1.07</td>
<td>0.049</td>
<td>0.61</td>
</tr>
<tr>
<td>Meridiani</td>
<td>2.1°S</td>
<td>61.1°W</td>
<td>5</td>
<td>0.028</td>
<td>0.18</td>
<td>0.013</td>
<td>0.064</td>
<td>0.47</td>
<td>0.026</td>
<td>0.50</td>
</tr>
<tr>
<td>Isidis</td>
<td>4.3°N</td>
<td>272.0°W</td>
<td>14</td>
<td>0.077</td>
<td>0.59</td>
<td>0.030</td>
<td>0.100</td>
<td>0.14</td>
<td>0.047</td>
<td>0.63</td>
</tr>
<tr>
<td>Athabasca</td>
<td>9.0°N</td>
<td>205.1°W</td>
<td>13</td>
<td>0.234</td>
<td>0.86</td>
<td>0.082</td>
<td>0.198</td>
<td>0.51</td>
<td>0.079</td>
<td>1.04</td>
</tr>
<tr>
<td>Melas</td>
<td>9.1°S</td>
<td>76.4°W</td>
<td>12</td>
<td>0.033</td>
<td>0.28</td>
<td>0.014</td>
<td>0.082</td>
<td>0.53</td>
<td>0.033</td>
<td>0.44</td>
</tr>
<tr>
<td>Eos</td>
<td>13.3°S</td>
<td>41.4°W</td>
<td>17</td>
<td>0.041</td>
<td>0.26</td>
<td>0.018</td>
<td>0.110</td>
<td>0.63</td>
<td>0.042</td>
<td>0.44</td>
</tr>
<tr>
<td>Region A</td>
<td>68.0°N</td>
<td>100.0°W</td>
<td>17</td>
<td>0.047</td>
<td>0.06</td>
<td>0.023</td>
<td>0.109</td>
<td>0.13</td>
<td>0.051</td>
<td>0.45</td>
</tr>
<tr>
<td>Region B</td>
<td>67.5°N</td>
<td>230.0°W</td>
<td>17</td>
<td>0.093</td>
<td>0.24</td>
<td>0.041</td>
<td>0.173</td>
<td>0.33</td>
<td>0.074</td>
<td>0.56</td>
</tr>
<tr>
<td>Region C</td>
<td>70.0°N</td>
<td>280.0°W</td>
<td>17</td>
<td>0.011</td>
<td>0.01</td>
<td>0.006</td>
<td>0.014</td>
<td>0.00</td>
<td>0.007</td>
<td>0.78</td>
</tr>
<tr>
<td>Region D</td>
<td>70.0°N</td>
<td>117.0°W</td>
<td>17</td>
<td>0.036</td>
<td>0.12</td>
<td>0.017</td>
<td>0.066</td>
<td>0.00</td>
<td>0.033</td>
<td>0.52</td>
</tr>
</tbody>
</table>

We observe that Region B has cross sections in family with VL2, and with the proposed Athabasca MER site, both with <18% rocks. With this, our main argument for confirming the 18% rock coverage at Region B remains the geomorphic analogy between Region B and VL2.

Ice cover issue: All Region B data were acquired at Ls~18°, at the end of northern winter. Aharonson et al [9] suggest that at the Region B latitude the CO$_2$ ice accumulation would only be of order 20 or 30 cm at that time. As cold CO$_2$ ice is both transparent and has real $\varepsilon$~2.3, this should not greatly affect the analyses. However, unanalyzed GVLA data from 1999 during northern summer will certainly improve the fits (Fig.1). Future work will refine layer modeling of ice over rough surface. In particular one might assume a smoothing of the annealing ice surface relative to the rock covered ‘dry’ surface and ask how a rougher interface under an 8-wavelength-thick layer of transparent material with a less rough surface influences diffuse backscatter at ~70° incidence angles. A tractable problem upon which we hope to report on at the meeting.

REGIONAL POLAR GLACIATION IN THE HESPERIAN PERIOD OF THE HISTORY OF MARS: THE SOUTH CIRCUMPOLAR DORSA ARGENTEA FORMATION AS AN ANCIENT ICE SHEET REMNANT. J. W. Head¹ and D. R. Marchant², ¹Dept. Geol. Sci., Brown Univ., Providence, RI 02912 USA (james_head@brown.edu), ²Dept. Earth Sci., Boston Univ., Boston, MA 02215 USA.

Introduction and Background: Polar regions represent cold traps for planetary volatiles and analysis of these areas permits an assessment of the amounts and types of volatiles, their stability and mobility, and the geological record of climate change. Present polar deposits on Mars consist of a thin residual ice unit (Api) overlying a thick sequence of layered deposits (Apl), and are of Late Amazonian age [1,2]. The individual layers in the current deposits are thought to be related to variations in orbital parameters [3], which cause changes in insolation and climate, and corresponding variations in dust and volatile stability, mobility, transport and deposition [e.g., 4,5]. Recent analysis of the history of orbital parameters has shown that the current martian climate is likely to be anomalous, and that Mars may have spent much of its history at considerably higher obliquity than its present value [3]. Indeed, glacial deposits at mid-latitudes [6,7] and huge tropical mountain glaciers [8-10] dating from earlier in the Amazonian are testimony to the mobility of polar ice and its transport and deposition equatorward during periods of higher mean obliquity [11].

More uncertain, however, is the nature of the climate of Mars prior to the Amazonian period. Do the polar deposits wax and wane during these climate changes or do they completely disappear? What evidence is there for the earlier climate history of Mars? It has been hypothesized on the basis of the nature and distribution of valley networks that Noachian Mars may have been "warm and wet" [12], that a global cryosphere developed subsequently [13] and that the global cryosphere was breached in numerous places and huge quantities of subsurface water were released into the northern lowlands during the Late Hesperian [14]. What was the nature of polar deposits during this earlier period? What was the volatile inventory and where was surface and near-surface water sequestered? Was the groundwater system leading to the outflow channels fed from basal melting of south polar ice deposits [13]? Did the outflow channel deposits lead to temporary but fundamental changes in climate [15]? Did the bodies of water formed by the outflow channels immediately migrate to the poles to produce polar ice sheets [16]? To address these and related questions we have been studying the south circumpolar Dorsa Argentean Formation, a unit mapped as Hesperian in age and displaying evidence of features related to glaciation. Here we summarize the nature of the DAF and implications for Hesperian polar environments and history.

The Dorsa Argentean Formation: The set of Hesperian-aged south circumpolar deposits represented by the Dorsa Argentean Formation (DAF) [e.g., 1,17,18] has been interpreted to be a volatile-rich polar deposit representing more than twice the area of the present Amazonian-aged layered terrain and residual polar ice, which it currently underlies. This huge polar ice-related deposit makes up about 2% of the surface of Mars and has undergone significant evolution since its emplacement. The deposit characteristics (e.g., smooth, pitted and etched deposits, pedestal craters, sinuous ridges interpreted as eskers, fluvial channels around the margins, marginal ponds and lakes, etc.) indicate that the DAF contained significant quantities of water ice, and that it represented a circumpolar ice sheet that subsequently underwent meltback and liquid water sub-ice sheet drainage, ponding in adjacent valleys, and ultimately draining, through surface subaerial channels, down into the Argyre basin more than 1000 km away. Among the most critical questions is the cause of the melting, the fate of the meltwater, and whether surface water reentered the groundwater system to recharge the global hydrologic system [e.g., 13]. Sources of melting include: 1) top-down heating: from general atmospheric evolution and global climate change, or from orbital-parameter-induced polar insulation changes, 2) bottom-up heating: from a general increase in geothermal flux, from specific intrusions or extrusions leading to magma-ice-contact melting, or: from increased ice thickness and load-induced basal melting.

Particularly striking examples of magma-ice-contact melting are candidate subglacial volcanoes within the DAF [19]. Seventeen anomalous mountains (originally mapped by [1], and now called Sisyphi Montes) form an unusual cluster in the central part of the DAF. The mountains occur over a large area, have separation distances of ~175 km, are typically 30-40 km in diameter, and ~1000-1500 m high, with their bases near ~1200 m elevation. Many members of this population are located on or adjacent to a 660 km long line extending toward the south pole. On the basis of their morphology, distinctiveness, alignment and isolation relative to other landforms, these features have been interpreted to be predominantly of volcanic origin [19]. A significant number of these features show unusual shapes, including flat tops and flat tops with a summit cone. Several of the mountains display sinuous channels around their margins and bases. These and other characteristics are interpreted to mean [19] that many of the mountains represented volcanoes that had erupted subglacially beneath the DAF (tuyas), and that the meltwater products could be traced from these regions to the margins of the deposit where they drained along eskers and channels into adjacent lakes or distant basins. The topography of the mountains suggested that the ice sheet averaged at least 1.4 km thick at the time of the eruptions.

Elsewhere in the Dorsa Argentean Formation, the topography is disrupted by a series of large irregular depressions (Cavi Angusti) whose origin has been attributed to eolian deflation and subglacial melting [e.g., 20,
and references therein]. Analysis of the largest of these depressions (~ 50 x 100 km in diameter and up to about 1500 m deep) shows terraced interiors, centrally located equidimensional and elongated structures interpreted to be edifices, and associated lava flow-like structures. An equidimensional mountain ~12 km in diameter and ~770 m high is centrally located within the basin, has anomalously steep sides and a flat top, and is perched on a low platform with lobate edges, extending about 2.5 km away from the edifice base in all directions. A lobate flow-like feature, 30 km long and 14 km wide, with clear terminal scarps, extends away from the base of the mountain toward the north, parallel with the elongate trend of the basin. The elongated structure is a ridge located to the northwest of the central mountain, is elongated in the direction of the long axis of the basin, has a similar height, and also rests on a platform. Together, these edifices and lobate structures are interpreted to be volcanic edifices and associated lava flows. Their central location in the depression in the Dorsa Argentea Formation strongly suggests that these features represent subglacial eruptions, and that their formation is directly related to the presence of the large depression, which is interpreted to be due to volcanic/ice interactions and melting [e.g., 20]. Indeed, volume estimates and heat transfer calculations [20] are consistent with such a mechanism involving a combination of intrusion and subglacial extrusion similar to that observed in Icelandic subglacial eruptions and meltwater generation.

Regional topography and ice-sheet geometry strongly suggest that any meltwater generated would drain to the north into the adjacent low areas. Evidence that this occurred includes an outlet and broad sinuous channel at the northern end of the largest depression, an unusual set of features interpreted to be a lake margin environment [e.g., 21] at the edge of the DAF less than ~150 km to the north of the basin, and a 300 x 800 km depression interpreted to be the site of a lake, which itself drains to the north into the Argyre basin [18]. Seven additional basins in Cavi Angusti contain mountains and ridges, usually centrally located, which are also interpreted to be the remnants of subglacial eruptions, formation of englacial lakes, and subsequent meltwater drainage to the north [20]. In summary, new spacecraft data support the interpretation that a significant part of the geomorphology of the Cavi Angusti region of the DAF is plausibly interpreted to be due to volcanic/ice interactions [20], an interpretation originally proposed by Howard [22] using low-resolution image data.

Furthermore, the eastern part of the volatile-rich DAF shows evidence of meltback, drainage and ponding of meltwater [23] adjacent to the region of interpreted subglacial volcanoes [e.g., 19]. Channels leading from the margins of the DAF enter nearby craters, and channels connecting the craters provide evidence for extensive crater flooding, ponding and filling, overtopping, downcutting, and further drainage through a series of craters into the Prometheus Basin, over a distance of ~600 km and involving a total vertical drop of ~800 m. Topographic evidence indicates that water filled some craters to depths of at least 200 m, and possibly up to 600 m, with minimum volumes of 10^{12} m^3. Along the central and western margins of the DAF, five sinuous valleys begin near the DAF edge and are carved into surrounding Noachian cratered terrain, extending for distances of up to 1600 km before emptying into the Argyre Basin, ~1-3 km below their starting elevations [18, 24]. The extension of these valleys into the DAF can be traced for hundreds of km due to the presence of aligned linear pits and basins and some preserved esker-like features on their floors [24]. The directions lead to the regions of Sisyphi Montes, the collection of isolated and aligned mountain features interpreted to be subglacial volcanoes [19].

Summary and Conclusions: The margins of the huge circum-south polar Dorsa Argentea Formation, interpreted to be an Hesperian-aged polar ice sheet, show evidence of extensive eskers [e.g., 18], marginal lakes [e.g., 21, 23], and drainage channels extending from the DAF margins for hundreds of km into surrounding depressions such as the Argyre basin [24]; these features provide evidence that recharge of the global aquifer is very likely to have taken place during this time. Although polythermal glaciation, due to the accumulation of polar ice deposits to thicknesses in excess of 3-4 kilometers (thus raising the melting geotherm into the base of the thickest part of the ice, permitting basal melting [13]) or higher global geothermal gradients in earlier martian history, cannot be ruled out, there is compelling evidence that a significant part of the meltwater is related to subglacial volcano-ice interactions. We believe that a significant portion of the volatiles may remain in the deposit, and thus that they were removed from the active hydrologic system. Therefore, the Dorsa Argentea Formation appears to represent an accessible polar climate record dating from early Mars history.

CHRONOS: A journey through martian history.  M. H. Hecht\(^1\) on behalf of the Chronos Team. \(^1\)Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Dr., Pasadena, CA 91109. (mhecht@jpl.nasa.gov.)

**Introduction:** In analogy with Earth’s ice sheets, the key to martian climate history lies below the surface, recorded in pristine stratigraphy that has not been scoured by modern surface processes and is not readily visible from orbit. Following methods established in Greenland and Antarctica, Chronos proposes to tell the story of martian climate history as a simpler analog to that of Earth’s, devoid of the influences of oceans, tectonics, and biology.

Four years ago, for the 2007 Scout opportunity, some of us proposed the CryoScout mission, designed to drill tens of meters through the polar ice of Mars. From the lessons of that proposal we developed and tested a new, low power approach to drilling in ice with up to 50% dust content.

**Scientific Objectives:** The largest visible reservoir of water on Mars, the North PLD is the only known unmodified and accessible record of recent Mars climate history. Chronos proposes to examine the conditions under which the north PLD were laid down over the past million years, thereby determining the recent climate history of Mars. Chronos’ log of images and compositional data would reflect the influence of meteorology, depositional episodes (volcanic, impact, dust storms), and planetary orbital/axial modulation.

By studying stratigraphy and sedimentology Chronos proposes to explore the historical relationship between meteorology and morphology. As secondary objectives Chronos proposes to characterize present-day polar surface/atmosphere interactions and to take advantage of the ideal geophysics platform offered by the polar cap to determine the geothermal gradient of Mars and explore its inner structure with seismometry.

**The Stratigraphic Record:** The PLD are believed to preserve a stratigraphic record of climate change over millions of years, modulated by quasi-periodic changes in the planet’s orbit and obliquity [1-3]. During accumulation periods, layers of water ice and dust are deposited and form chronological units characteristic of global and local conditions at the time of deposition [e.g., 4-7]. During some periods in their history the PLD have experienced net ablation, causing discontinuities in the record by removing earlier layers [8].

Chronos proposes to resolve uncertainty about the characteristic timescale represented by individual layers. Early estimates based on Viking-era imagery assumed that the ~30 m layer packets correspond to 10^5-10^6 year periodicity in Mars’ obliquity, suggesting that annual layers may be as thin as 0.01 mm [9]. Since then, high-resolution MOC images [10] combined with MOLA topography have offered new insights into layer formation (Fig. 1), suggesting that the layers may have formed on 51,000 year scales corresponding to precession of perihelion [11-12]. This rate would imply annual layers on average ~0.5 mm thick, consistent with resurfacing rates derived by Herkenhoff and Plaut [13] from the lack of craters on the North PLD (though resurfacing may be due to annual cycles of ablation and deposition).

The Chronos objective is to reconstruct a polar chronology from the observed stratigraphy, the chemical and isotopic data, and orbital models. In Greenland ice cores this method has achieved an absolute accuracy of 1% in the top 2500 m [14-15]

**Proposed measurements:** Chronos proposes to drill from 10-75 m below the surface of the ice and to perform the following measurements and analyses:
- Image visible stratigraphy as an indicator of ice and dust deposition rates, resolving ~10^-3 m features. Counting layers will be the simplest approach to dating the polar stratigraphy.
- Measure the concentration of dust as a function of depth and constrain its size distribution.
- Measure isotopic ratios as a direct indicator of ice accessibility and sublimation or condensation rates, and an indirect indicator of temperature and other primary meteorological parameters. These ratios will also be sensitive to the solar cycle, which affects the upper atmosphere ablation rate, possibly offering an alternative clock for the chronological record.

![Fig. 1: Estimated depth for a sequence of layers in the PLD; depth values are from MOLA data. These are not the exact layers Chronos will traverse but are typical of the overall structure.](image)
• As a function of depth, identify, quantify and characterize the dissolved inorganic chemical components in the meltwater resulting from the deposited martian dust. Monitor the pH of the meltwater to constrain the total atmospheric pressure.

• Image the surface topography and morphology of the ice and ascertain the surface reflectance in order to constrain the thermal budget.

• On the surface, Measure humidity, T, P, and wind to determine net flux of water to/from surface. Constrain seasonal deposition & removal rate.

• Determine the geothermal gradient in the PLD

• Constrain the internal structure of the planet from response of the ice sheet to seismic activity

**Technology:** Chronos drills passively create a melt front at the nose, then pump melt water to the lander deck (Fig. 2). Pumping meltwater to the lander not only provides an effective method for sample collection but also minimizes thermal contact between the drill and the ice, saving significant amounts of power. A 7.5 cm diameter drill, with the application of 200–400 W, can descend at speeds of 20–45 cm/hr in 163 K ice. The drill is suspended from a tether, which provides down-hole power and data transfer and keeps the meltwater liquid all the way to the surface.

Each drill carries a downhole camera to record visible stratigraphy, acquiring full-color stereo images at 10^{-5} m per pixel, sufficient for observing annual layers. The longer of two drills also determines the geothermal gradient with a temperature sensor, while the shorter carries a miniature seismometer to explore the inner structure of the planet. On the lander deck an isotopic laser spectrometer measures variations in relative hydrogen- and oxygen-isotope abundance in the meltwater, reflecting source and climate conditions under which the ice was deposited. A companion instrument uses electrochemical sensors to characterize embedded dust in the meltwater, determining the salt composition and abundance. The surface payload includes a panoramic imager and a meteorology station that measures pressure, temperature, humidity, and wind. Together, they characterize surface dynamics to determine both thermal and water balance at the ice/atmosphere boundary [16].

**Acknowledgement:** This drill described here was developed at the Jet Propulsion Laboratory, California Institute of Technology, with support from NASA’s PIDDP, ASTID, and ASTEP programs.

**References:**

UNDERSTANDING AMBIGUITIES IN BACKSCATTERED ORBITAL RADAR SOUNDING DATA FROM THE MARTIAN POLAR LAYERED DEPOSITS USING FINITE DIFFERENCE TIME DOMAIN SIMULATIONS. E. Heggy, S. M. Clifford, D. Card, Lunar and Planetary Institute, Houston, Texas, USA (heggy@lpi.usra.edu and clifford@lpi.usra.edu), University of Manitoba, Department of Geological Sciences, Canada,

The MARSIS orbital radar sounding data has provided our first insights into the internal structure, stratigraphy, and basal topography of the polar-layered deposits (PLD) [1]. In order to evaluate the impact of the various geoelectrical and structural characteristics of the deposits on the sounding data obtained by MARSIS and anticipated from SHARAD, we have constructed representative geoelectrical profiles of the Martian PLD using new laboratory dielectric measurements of ice and ice-dust mixtures over the frequency range of 1-20 MHz [2]. These models address the impact of variable dust content, dust composition, layering, and the potential presence of gas hydrates or thin films of super-cooled water) that might be the most plausible explanations for the observed reduction in reflectivity associated with the surface of the SPLD, we considered randomly oriented 150-m wide fractures (with a horizontal density of 1/2250 m²) with dips ranging from 30° – 80°, that penetrated to a vertical depth of 150 m. These characteristics where investigated only to provides a first order understanding of the effect of surface fractures on signal dispersion and depolarization. A more comprehensive range of fracture characteristics is currently being investigated. Our initial results indicate that fracture geometry and orientation can significantly affect surface reflectivity at 20 MHz, but that the effect is reduced at 2 MHz.

In order to simulate the radar echo for the layered stratigraphy, we used the FDTD algorithm, which is very general in terms of material electrical properties, geometry and frequency definition [3]. We evaluated the scattered electric field in two cross polarizations at the surface of our geoelectrical models. Simulations were performed in the time domain to observe reflections at each interface. In FDTD modeling, the material properties and geometry are constructed using the Yee cell method [4]. For each frequency, each cell is characterized by the complex permittivity of the investigated layer. We used the Perfect Matching Layer (PML) algorithm as boundary condition to reduce signal reflections at the column boundaries to decrease simulation noise.

For the 2 MHz simulations that correspond to the MARSIS case, the necessary dimensions of the Yee cell are 5x5x5 m in order to insure sufficient accuracy and stability in the FDTD calculations. To describe the roughness at the layer interfaces, and the scatterers in the first layer, we use a local grid with smaller cells of 1 m. The radar pulse emitter and receiver are placed at the same point above the surface. The vertically emitted pulse is simulated as a plane wave with maximum amplitude of 1 V/m. The emitted waveform is a vertically polarized (in the x direction) and modulated Gaussian, with a central frequency of 2 MHz and 2 MHz bandwidth. The receiver can measure the backscattered echoes in the Ex and Ey cross polarizations. For SHARAD 20-MHz simulations, the size of the Yee cell is reduced to 1 m.

In order to evaluate the effect of the first layer on electromagnetic attenuation, we changed its complex permittivity according to the measured range, and then observed the effect on the radar echo. To evaluate the volume scattering effect, particularly their potential contribution to the observed reduction in reflectivity associated with the surface of the SPLD, we considered randomly oriented 150-m wide fractures (with a horizontal density of 1/2250 m²) with dips ranging from 30° – 80°, that penetrated to a vertical depth of 150 m. These characteristics where investigated only to provides a first order understanding of the effect of surface fractures on signal dispersion and depolarization. A more comprehensive range of fracture characteristics is currently being investigated. Our initial results indicate that fracture geometry and orientation can significantly affect surface reflectivity at 20 MHz, but that the effect is reduced at 2 MHz.

At the conference, the results of these simulations will be summarized and compared with the actual MARSIS SPLD data, identifying what we believe to be the most plausible explanations for the observed surface and basal reflectivity of the deposits, as well as the range of characteristics capable of explaining the differences between the visual- and radar-determined stratigraphy.
ON THE DIELECTRIC PROPERTIES OF DUST AND ICE-DUST MIXTURES: EXPERIMENTAL CHARACTERIZATION OF THE MARTIAN POLAR LAYERED DEPOSITS ANALOG MATERIALS. E. Heggy1, S. M. Clifford1, A. Younsi2, J.L. Miane2, R. Carley3, R.V. Morris4, 1Lunar and Planetary Institute, Houston, Texas, USA (heggy@lpi.usra.edu and clifford@lpi.usra.edu), 2Laboratoire PIOM-ENSCPB, Pessac, France; 3University of Cambridge, Cambridge, UK; 4NASA Johnson Space Center, Houston, Texas, USA.

Recent data from the MARSIS orbital radar sounder on Mars Express has provided the first insights into the electromagnetic properties of the Martian polar layered deposits (PLD). Among the most important information that can be derived from the MARSIS polar data are the thickness, depth and geometry of internal layers. A necessary requirement for the accurate interpretation of these data is knowledge of the appropriate dielectric properties of the polar ice as a function of dust concentration, composition, temperature, and radar sounding frequency. To address this need, we have conducted laboratory measurements of the electromagnetic properties of several dry Mars analog soils that match the surface compositions inferred from the TES [1] and OMEGA [2] IR spectral data. To investigate how the presence of such material affects the dielectric properties of the polar ice, we have also conducted measurements of ice-dust mixtures, covering the frequency range from 1 MHz to 3 GHz. Measurements were performed as a function of density, temperature, dust content and dust composition. Measurements were performed on two types of samples: (1) dry soils -- both natural and synthetic -- consisting of ground basalts and various mixtures that included components consisting of hematite, magnetite and maghemite to pure silica sand; and (2) ice-dust mixtures with various concentrations of the dry soil analogs prepared in (1). These measurements are being compiled to construct more realistic geo-electrical models of the PLD and better interpret the MARSIS data.

Introduction: The Mars Advanced Radar for Subsurface and Ionosphere Sounding (MARSIS), which operates in the frequency band of ~1 to 5 MHz, represents the first attempt to explore the Martian deep subsurface, with a particular emphasis on the detection of subsurface water [4]. MARSIS radargrams of the PLD have revealed significant internal layering, but whose relationship to the visible layering observed in the polar scarps and troughs is unknown. Early analysis of these data suggests that their accurate interpretation is strongly dependent on our understanding of the electromagnetic properties of the deposits [5]. On Earth, this type of knowledge is often acquired with the aid of complementary electromagnetic sounding techniques – such as Transient Electromagnetic Methods (TEM) and resistivity measurements. These techniques provide an independent assessment of soil electrical conductivity which, in turn, affects radar signal attenuation and penetration depth. While this knowledge can significantly reduce the ambiguities in radar data interpretation, such an approach is not yet possible for Mars. Instead, we must rely on laboratory measurements of analog materials to help constrain the plausible range of dielectric properties that define the propagation and reflection characteristics of the PLD. The measurements presented here span the range (1 MHz - 3 GHz) covered by MARSIS (1 – 5 MHz), the SHARAD (20 MHz) sounder onboard the Mars Reconnaissance Orbiter, and the WISDOM GPR (0.5 – 3 GHz) that will fly on ESA’s 2011 ExoMars Rover [6].

Experimental setup: The soil/dust samples were prepared by grinding each of the mineral components into a fine (50 μm grain) homogenous powder, which were then mixed in various amounts to form the simulants. Sample density and porosity were controlled using a hydraulic press to compact the powders into pellets having equal masses. Samples were then dried in a vacuum-oven for 48 hours in order to remove residual moisture that can affect the measured complex dielectric constant. To prepare the ice-dust samples, dust was mixed with different amounts of distilled water to form a homogenous liquid, which was then poured into the dielectric cell and frozen (while being agitated to maintain mixture uniformity). Two cells were used that were specially designed for measuring the dielectric properties of low loss materials (e.g. teflon and ice). The first is an open coaxial cell used to measure the dielectric constant of loose powder materials. The second is a reflection cell that is used to measure ice-dust mixtures. Both measurements cells are connected to an impedance analyzer that swept the frequency and conducted measurements over the range of 1 MHz - 3 GHz. The two cells were placed in an environmental chamber which permitted control and variation of both temperature and pressure conditions after the samples achieved thermal equilibrium. The analyzer is connected to a central command unit to extract data and calculate in real-time the real and imaginary parts of the complex dielectric constant. Measurements were then made over the full frequency range for each sample as a function of temperature, density, dust content and composition.

Measurement results: Figure 1 shows the real and imaginary parts of the dielectric constant, as a function of the samples density, for a typical loose, dry basaltic soil sample, containing ~14% of magnetite [7],
that was collected near Craters of the Moon National Monument (Idaho, USA).

![Figure 1: The real (top) and imaginary (bottom) part of the dielectric constant for a loose, dry basaltic soil as a function of density and frequency, at room temperature.](image)

The lab measurements show a clear increase in both the real and imaginary parts of the dielectric constant with increased density. However, their spectral behavior is different: the real part tends to decrease with the frequency, while the imaginary part shows a very slight increase. This spectral behaviour is consistent as a function of density but varies with composition. As for temperature, our current equipment has been calibrated to operate at a minimum of –80°C with errors in the complex permittivity below 5%. In Table 1, dielectric measurements of the same loose, dry basaltic soil as in Fig. 1 are shown over the temperature range of -20°C to -70°C at a density of 1.9 g/cm³. Over this 50°C temperature range, the measurements indicate that the influence of temperature is not as significant as density on the real and complex parts of the dielectric constant.

![Figure 2: The real part of the dielectric constant for an ice-basalt mixture.](image)

At 2-20 MHz, the real part of the dielectric constant ranges from 3.1 for pure ice to 4.8 for an ice sample with a 75% basaltic dust content. Measurement of the imaginary part of the dielectric constant, for this and other ice-dust mixtures, will be present at the conference. These measurements are expected to greatly assist in assessing the attenuation and maximum sounding depth of MARSIS, as well as the surface and internal layer reflection coefficient for the polar deposit materials.

**References:**

IMAGING THE POLAR REGIONS OF MARS WITH HIRISE. K. Herkenhoff1, S. Byrne2, C. Hansen3, V. Gulick4, and A. McEwen2, 1U. S. Geological Survey Astrogeology Team, Flagstaff, Arizona, USA (herkenhoff@usgs.gov), 2University of Arizona, Tucson, USA, 3Jet Propulsion Laboratory, Caltech, Pasadena, California, USA, 4SETI Institute, Moffett Field, California, USA.

Introduction: The Mars Reconnaissance Orbiter (MRO) was launched in August 2005 and entered Mars orbit in March 2006. Orbit insertion was followed by 6 months of aerobraking to achieve a low (300 km), 2-hour Mars orbit. Aerobraking should be completed by early September, followed by 2 months of transitional orbits and solar conjunction on October 23. The MRO payload includes the HiRISE camera [1], which features a 0.5 m diameter primary mirror, 12 m effective focal length, and a focal plane system that can acquire images containing up to 28 Gb of data in as little as 6 seconds. HiRISE will provide detailed images (0.25 to 1.3 m/pixel) covering ~1% of the martian surface during the 2-year Primary Science Phase (PSP) beginning November 2006. The images may include color data (green, red, near-IR) over 20% of the field of view. A top priority is to acquire ~1000 stereo pairs and apply precision geometric corrections to enable topographic measurements to better than 25 cm precision. We expect to return at least 12 Tb of HiRISE data during the 2-year PSP and use pixel binning, conversion from 14 to 8 bit values, and a lossless compression system to increase coverage. HiRISE images are acquired via 14 CCD detectors, each with 2 output channels, and with multiple choices for pixel binning (none, 2x2, 3x3, 4x4, 8x8, 16x16) and number of Time Delay Integration lines (8, 32, 64, or 128). HiRISE will support Mars exploration by locating and characterizing past, present, and future landing sites, crash sites, and rover traverses. We will investigate cratering, volcanism, tectonism, hydrology, sedimentation, stratigraphy, aeolian processes, mass wasting, landscape evolution, seasonal processes, climate change, spectrophotometry, glacial/periglacial processes, polar geology, and other science themes.

HiRISE Polar Observations: MRO’s orbit will have an inclination of 93 degrees during PSP, allowing frequent imaging of the polar regions. MRO polar campaigns are being planned jointly among the MRO science teams. HiRISE polar science objectives are divided into two themes: Polar Geology and Seasonal Processes.

Polar Geology. The deposits at both Martian poles can naturally be divided into the polar layered deposits and the much thinner residual ice caps which partially cover them. Several important questions related to the residual ices are: Are the residual caps currently accumulating or losing material? How much exchangeable carbon dioxide and water is locked up within these deposits? How stable are these deposits over timescales of centuries to millennia? How do the residual caps control the formation of the polar layered deposits? The relationship between layers within the polar layered deposits and the climate during their formation remains uncertain. The dust/ice ratio and how it varies from layer to layer is currently unknown; some layers may also contain other more minor components such as volcanic fallout and impact debris. Identification of these materials would help our understanding of recent Martian history significantly. HiRISE is capable of observing Martian layers at sub-meter resolution. Such observations can be used to determine whether these layered deposits contain even thinner layers than those that are resolved by MOC. An important goal of Mars polar science is the association of individual layers with independently derived Martian orbital changes. One of the more puzzling questions regarding the Martian polar regions is why the southern polar layered deposits appear to be so much older than those in the north. The difference in surface age is inferred by counting craters, and there are many more craters on the southern layered deposits than on the northern ones.

Changes in the shape and size of features within the southern residual carbon dioxide ice will be observable over timescales substantially less than 1 year. As indicated in Fig. 1, scarps in the residual cap retreat about 5 meters every Martian year. HiRISE will allow us to monitor these changes over many years. The northern residual ice surface has an unusual texture at the limit of MOC resolution. HiRISE observations will allow characterization of these small-scale landforms and their evolution rates. Layers less than 1 meter thick are visible on the sides of mesas in the southern residual ice. These layers can be imaged by HiRISE and may hold the key to Mars’ climatic history over the past few centuries.

High-resolution stereo imaging of exposed sequences of polar layered deposits will allow measurements of layer properties such as thickness and bedding attitudes. The layering is best observed in the summer, when seasonal frost has evaporated and the sun is still high enough to illuminate the polar region. Stereo images will be acquired as close in time as possible, so that differences in illumination and frost distribution are minimized. Observation of small craters on the
layered deposits will allow us to put a more accurate constraint on the timescales of deposition and erosion operating in that area as well as providing information about the target material and crater modification processes. HiRISE views of how these crater shapes have relaxed and where the icy layers have broken in a brittle way will give us a better idea of the strength of the layered deposit material and whether it has flowed significantly.

Observing the relationship between changes in the residual ices and the current climate will allow us to better understand how the layered deposits were formed. This information can be used as a ‘Rosetta stone’ for interpreting the climatic record within the layers themselves.

**Seasonal Processes.** Mars’ seasonal polar caps are composed primarily of CO$_2$ frost. To study seasonal processes we will image the caps as they wax and wane, observing both large scale effects on Mars as well as the local details of the sublimation and condensation processes. By learning about current processes on a local scale we can learn more about how to interpret the geological record of climate changes on Mars. Mars Observer Camera (MOC) images from the Mars Global Surveyor spacecraft have shown an astonishing array of exotic landcapes as the southern seasonal cap sublimates, including spots, “spiders”, and fans. A region we plan to investigate near the south pole has been called the “cryptic” terrain because it seems to stay quite cold even after the disappearance of bright frost.

One example of many phenomena we would like to observe is the evolution of a “spot” to a “fan” as the seasonal cap retreats in the spring. Is the darker material very fine and does it get blown across the surface of the brighter surrounding ice to form a fan? Or is the darker material lofted by Triton-like plumes such as those observed by Voyager 2? The high resolution and high signal-to-noise ratio of the HiRISE images along with stereo coverage will give us our best ever view of these unearthly terrains.

**Data Processing and Release:** We will calibrate, reproject, and mosaic the 28 HiRISE channels into Reduced Data Records (RDRs) and also produce Digital Elevation Models and special products for release via the Planetary Data System (PDS). The HiRISE group in Tucson is a PDS subnode, and we plan to release RDRs (along with the raw data) as soon as they have been completed and validated. An Internet web site (HiWeb) will enable anyone in the world to suggest HiRISE targets on Mars and to easily locate, view, and download HiRISE data products.

**HiRISE Education and Public Outreach:** The HiRISE team is planning an innovative education and public outreach program with a variety of formal and informal educational activities. These include educator workshops, large-scale displays of HiRISE images at museums and planetariums, and opportunities for students, scientists, and the general public to suggest HiRISE image observations and to participate in data analysis via the internet. The centerpiece of the E/PO program is HiRISE’s interactive website called HiWeb ([http://marsweb.nasa.nasa.gov/hirise](http://marsweb.nasa.nasa.gov/hirise)) [2,3]. HiRISE’s public website will provide tools to suggest target locations for HiRISE imaging. HiWeb will provide interactive viewing and analysis of HiRISE images in context with other available Mars data. Web events, involving participation by team members, will inform students and interested members of the Mars science community and the public of HiRISE capabilities and science goals and provide support for submitting good image suggestions. Curriculum modules and activities ([http://hirise.seti.org/epo](http://hirise.seti.org/epo)) will focus on Mars geology, the image suggestion process and working with digital image data. We will also provide online opportunities for student and public participation in data analysis to create databases of geologic features (gullies, boulders, craters, wind streaks, etc.) present in the HiRISE images. Educator workshops will be held each year at or near the institution of HiRISE team members. Workshop background materials and instructions for all hands-on activities will be placed on HiWeb to facilitate sharing of information with other educators and the general public.

**References:**


**Figure 1.** MOC images of 8 m-deep pit (dubbed “Swiss-cheese” feature) at 86.9°S, 353.3°E separated by two Martian years. M09/00609 (left, L$_{237}$°, 1.4 m/px) and R08/01050 (right, L$_{241}$°, 1.5 m/px); illumination from lower left. Gridline spacing 50 m; expansion of pit relative to fixed grid can be seen.
MARS POLAR GEOLOGIC NOMENCLATURE: WHAT ARE THE CAPS?  K. E. Herkenhoff¹, S. Byrne¹, and K. L. Tanaka¹, ¹U. S. Geological Survey, Astrogeology Team, Flagstaff, Arizona, USA (kherkenhoff@usgs.gov). ²Lunar and Planetary Laboratory, University of Arizona, Tucson, Arizona, USA.

Introduction: The polar caps of Mars were first observed over 300 years ago, and seasonal changes were noted soon thereafter [1]. Early observers recognized that the polar caps do not completely disappear in the summer, leading to the term “residual” cap [2]. Similarly, the polar caps that wax and wane annually have been called the “seasonal” caps. This nomenclature, based on the higher reflectance of the caps, has survived into the modern era of spacecraft observations, when it became clear that seasonal caps are composed of CO₂. Mariner 7 images showed some evidence for layered deposits in the south polar region, and the polar layered deposits were clearly seen in Mariner 9 images [3]. The Viking Orbiters found that while the north polar residual cap is composed of water ice, the south polar residual cap is covered by CO₂ throughout the year. These missions and currently operating spacecraft have also shown that the geology of the polar caps and layered deposits is complex [4-7], with significant differences between the poles. Still, the simple “polar cap” terminology defined above is relevant and useful today.

However, the nomenclature used to describe geologic and topographic features in Mars’ polar regions varies significantly in recent publications. For example, the term “cap” has been used to describe the frost and ice that covers the polar regions seasonally, and to describe the topographic domes at each pole (Figure 1). These inconsistencies in terminology can cause readers to become confused. The Mars polar science community should use a standard set of nomenclature; here, we propose Mars polar nomenclature that is consistent with previous geologic mapping [4, 5].

Mars Polar Geologic Nomenclature: Geographic names can also be used when referring to large-scale topographic features, such as the north polar dome that includes the polar layered deposits, residual polar ice cap, and other geologic units. This topographic dome should be called “Planum Boreum” rather than the “north polar cap” because the latter term implies that the polar layered deposits are part of the north polar residual ice cap. Similarly, the broad topographic dome that includes the south polar layered deposits and residual cap should be called “Planum Australe,” as the south polar layered deposits extend far beyond the residual cap. These geographic names have been approved by the IAU, and are therefore unlikely to change in the near future. While recent geologic maps subdivide polar geologic units in considerable detail [5, 6], we feel that the following generalized nomenclature will suffice for most publications that discuss the Martian polar regions.

North polar nomenclature: The following informal names are in stratigraphic order:
- Seasonal CO₂ frost/ice cap
- Residual (permanent) ice cap
- Polar layered deposits
- Northern plains

South polar nomenclature: The following informal names are in stratigraphic order:
- Seasonal CO₂ frost/ice cap
- Residual (permanent) ice cap
- Polar layered deposits
- Hesperian plains (Dorsa Argentea Formation)
- Ancient highlands

Figure 1 illustrates in a very simplified form the relationship between the residual ice caps, layered deposits, and underlying terrain. Seasonal CO₂ covers everything in this figure each winter. Information on the basement topography is just becoming available so the lower contacts shown here are speculative.

Summary: The Mars polar nomenclature outlined above is consistent with published geologic maps and the majority of other publications. It restricts the use of the term “cap” to avoid ambiguity, while allowing the (likely) possibility that the polar layered deposits are composed mostly of water ice. We hope that journal editors and referees will work to ensure that Mars polar nomenclature is standardized in this way.

Figure 1: Schematic geologic maps and cross-sections of Planum Australe (South pole, left) and Planum Boreum (north pole, right). Blue and white coloring represent polar layered deposits and residual ice [4,5,6,7]. Planum Boreum is a dome-shaped structure situated on a slight incline. Planum Australe is dome-shaped at its thickest location (profile A-A' above), but forms an extensive plateau further equatorward (profile B-B' above) which thickens to the west. In the region shown in profile A-A', Planum Australe buries part of Promethei Rupes, the edge of a large impact basin. The residual ice caps are draped over the highest portions of both Planum Boreum and Australe and are indicated on the cross-sections with a thick black line (the thickness of which does not represent the real thickness of the residual ice). The volume of both residual ice caps is inconsequential compared to that of the underlying layered deposits. All three plots are on the same scale (1500x5 Km), the vertical exaggeration is 100:1. Parallels and meridians are spaced every 5 and 30 degrees respectively on the map views, yellow lines indicate cross-section locations.
Introduction: The martian polar layered deposits (PLD) are probably the best source of information about the recent climate history of Mars [1-7], but their origin and the mechanisms of accumulation are still a mystery [8]. The polar layers are sedimentary deposits that most planetary scientists believe are composed of water ice and varying amounts of wind-blown dust [2-4], but their composition is poorly constrained [9]. Because climate changes are likely recorded as variations in composition or deposition/erosion rates between layers, the detailed stratigraphy of the PLD is of great interest. Layer thicknesses of ~10 to 50 m were observed in Viking Orbiter images of the north PLD by Blasius et al. [10], and Mars Orbiter Camera (MOC) images resolve layers with similar or lesser thicknesses in both polar regions [11]. In order to accurately determine the thickness of layers and interpret PLD stratigraphy and structure, the topography of exposures must be known. Previous studies have identified deformation in the PLD similar to that observed in terrestrial glaciers [12-14], but lack of detailed topography has hindered structural interpretations. At the Third International Mars Polar Conference, we reported photoclinometric results for the south polar layered deposits. Here we describe results of our continuing study to evaluate the topography, structure, and stratigraphy of the north PLD using photoclinometry on MOC images.

Approach: We used a 2-dimensional photoclinometric technique [15] constrained using simultaneously-acquired MOLA data. This technique is well suited to images taken at high latitudes when the surface was covered by seasonal frost and the solar elevation angle was low so that albedo variations and their effects are minimized and topographic modulation is emphasized (Fig. 1). The high density of MOLA data in the polar regions allows gridded topographic products to be generated at higher spatial resolution (~250 m/pixel) than is possible at lower latitudes. We introduce MOLA topography into the process in five ways [16]: 1) for planimetric control; 2) to precisely model surface and atmospheric reflectance/scattering; 3) to account for subtle variations in surface albedo; 4) as the starting solution for the photoclinometric model; 5) as the DEM base map on which the MOC NA high-resolution DEMs are mosaicked. The ability to discern the layers facing the sun in the photoclinometrically-derived digital elevation model (DEM) is at essentially the same spatial resolution as the original image.

Figure 1. (left) MOC image M2001450, (middle) MOLA shaded relief (noisy), and (right) digital elevation model derived using photoclinometry. Images are each 3 km wide; illumination from upper right.
Results: Figure 2 shows some topographic profiles across the layers shown in Figure 1. Noteworthy results from this DEM are that (1) individual layers visible in the MOC images are typically 20-40 m thick, (2) many layers are characterized by a raised “rim” or “lip”, (3) while some layers are continuous across the 3 km width of the image, others cannot be traced for more than about a kilometer using only the profiles, and (4) the layers have an apparent dip of about 1° to the east, but different layers dip at different angles, suggesting pinching and swelling of some layers. Further analysis of photoclinometric models of this and other PLD exposures will be presented at the conference.

References:

![Figure 2. Profiles from photoclinometric DEM. Profiles are 600 m apart. Red lines indicate preliminary stratigraphic correlations.](8027.pdf)
ARCTIC ANALOGUE SCIENCE AS PART OF AN INTEGRATED CANADIAN STRATEGY FOR MARS EXPLORATION  V. J. Hipkin, Berinstein A, Laurin D, Ouellet A, Lebeuf M, Osinski G.R., Léveillé R, Canadian Space Agency, 6767 Route de l’Aéroport, St-Hubert, QC J3Y 8Y9, Canada (victoria.hipkin@space.gc.ca)

Introduction:  Canada has had a long involvement in space science beginning with the launch of the Canadian satellite Alouette 1, in 1962. However the Canadian Space Agency's involvement in planetary exploration began only in the late 1990's with the contribution of the Thermal Plasma Analyzer (TPA) to the Japanese Nozomi mission. The Canadian Space Agency (CSA) is now contributing science payloads to the first NASA Scout mission, Phoenix, and NASA's Mars Science Laboratory, and is exploring strategies to integrate existing science and technology capabilities towards a robust planetary exploration program.

Underpinning a Canadian exploration strategy, the Canadian Arctic forms a natural asset. Its cold, dry desert environments provide an operational analogue for Mars exploration, and its relatively unweathered impact craters, permafrost regions and arctic springs provide sites where processes relevant to Mars polar science can be studied.

Current program elements: Three programs make up the Mars Exploration framework at CSA (Figure 1).

The Canadian Analogue Program, including the Canadian Analogue Research Network (CARN) [1], is providing opportunities for a new generation of planetary scientists to apply terrestrial expertise to planetary questions. In 2006, its first summer of operation, the program will support a range of science investigations (Table 1).

In parallel, a Mars Instrument Concept Studies opportunity is encouraging the design and development of instruments (Table 2). This is part of an ongoing instrument development program.

These two programs can be closely related. Science investigations at analogue sites provide important input to measurement and detection methodologies for planetary missions. Key discoveries from planetary missions depend on drawing conclusions from a limited set of measurements – analysis of data for the presence of a definitive “signature”. The identification of a signature for primitive life, for example, is a well known challenge that is being addressed through studies of terrestrial analogues [2]. The derivation of new signatures from analogue studies will allow for the design of highly optimized instruments and investigations for the next generation of missions.

The RIGID concept (Table 2) is an example where instrument science requirements are being derived with the help of analogue studies at the CARN Mars Arctic Research Station site.

Once instrument concepts are in development, operational methodologies for landed missions can be tested at analogue sites. This will expose scientists to the reality of operations in a challenging environment: weather conditions, autonomy, power and bandwidth limitations.

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Making up the triad, the CSA Space Technology Development Program supports research in core technologies as well as exploration infrastructure eg. development of drilling technologies based on Canadian mining and exploration geology expertise [3]. ISRU and operational methodologies for human exploration can be expected to become an increasing focus of the Canadian analogue program as the prospect of manned missions to the moon and Mars nears.
Future Opportunities: These three programs provide for grass roots development for future Canadian contributions to Mars Polar Science. An expectation is that the Canadian exploration program will mature to include planetary missions as well as support for international flagship missions. A CSA call for Mars mission concepts closed summer 2006, to enable teams to form and develop ideas. Development of a Canadian roadmap for exploration has begun and will continue in consultation with the Canadian and international communities.


Additional Information: Participation in the CSA workshop "Exploration Canada", Oct 2006, and in the 6th Canadian Space Exploration Workshop, June 2007 (following on to address the priorities of the Canadian scientific community in more detail), is warmly welcomed. Contact: Dr V. J.Hipkin, Program Scientist for Planetary Exploration, Canadian Space Agency. Email: Victoria.Hipkin@space.gc.ca

<table>
<thead>
<tr>
<th>PI</th>
<th>Title</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bernath</td>
<td>ACE-M: Atmospheric Chemistry Experiment for Mars</td>
</tr>
<tr>
<td>Cloutis</td>
<td>Mars Hyperspectral Imager</td>
</tr>
<tr>
<td>Daly</td>
<td>Microscopic Imager for In Situ Martian Geological Analysis</td>
</tr>
<tr>
<td>Herd</td>
<td>Borehole Gamma Ray Spectrometer for the subsurface Exploration of Mars</td>
</tr>
<tr>
<td>Kruzelecky</td>
<td>Integrated Sample Analysis Instrument Concept for Rover and Lander Platforms using Advanced Miniaturization and Fiber-Optic Technologies for VIS/IR Spectrometers.</td>
</tr>
<tr>
<td>Pollard</td>
<td>Resistivity Instrumentation for Ground Ice Detection (RIGID)</td>
</tr>
<tr>
<td>Prasad</td>
<td>Geophysical Sensors for Mars Exploration</td>
</tr>
<tr>
<td>Ward</td>
<td>Imaging Doppler Michelson Interferometers for Winds in the Martian Upper Atmosphere</td>
</tr>
</tbody>
</table>

Figure 1: Canadian Space Agency Mars Exploration Framework
AIRBORNE RADAR SOUNDING IN ANTARCTICA: ANALOGS FOR SHARAD DATA OVER THE POLAR LAYERED TERRAINS AND VALLEY GLACIERS OF MARS.  J.W. Holt1, M.E. Peters1, D.D. Blankenship1, H. A. Danque1, and S. Carter1, University of Texas Institute for Geophysics, The John A. and Katherine G. Jackson School of Geosciences, University of Texas, 4412 Spicewood Springs Rd., Bldg. 600, Austin, TX 78759, jack@ig.utexas.edu

Introduction: The SHAllow RADar (SHARAD) instrument on Mars Reconnaissance Orbiter [1, 2] will provide a new dimension to studies of glacial features on Mars that have been rather well characterized through optical and spectral imagery. It will also expand our ability to study internal features of the Polar Layered Deposits (PLDs) at high resolution relative to the successful soundings accomplished by MARSIS on Mars Express [3-5]. Antarctica contains a number of features analogous to those on Mars. The University of Texas Institute for Geophysics has been conducting airborne geophysical studies of Antarctica since 1990 [6, 7], including radar sounding with an instrument that is similar to SHARAD in several ways.

Given that gravity, magnetics and laser altimetry data are typically acquired simultaneously with our radar sounding, we have other constraints in our interpretations of Antarctic features; however, much can be learned from the radar data alone, especially when the appropriate analyses are performed to quantitatively interpret the data. Well-constrained analogs from Earth with multiple data sets should assist in better understanding radar results from Mars. Here we provide examples of data and analysis techniques that may help in planning for the acquisition and interpretation of SHARAD data from similar features on Mars.

Radar Comparison: SHARAD is an orbital, chirped radar operating at a 20 MHz center frequency (15 meters free-space wavelength) with 10 MHz bandwidth and 85 µs pulse duration [1, 2]. The current UTIG radar system is a coherent, chirped radar operating at a 60 MHz center frequency (5 meters free-space wavelength) with 15 MHz bandwidth, 8 kW transmit power, and 1 µs pulse duration. The electrical properties of rock and ice are not substantially different at these frequencies. The same features should appear quite similar in both radars, especially over the relatively flat polar ice deposits. Pulse-compression and synthetic aperture processing can be applied to both. An earlier UTIG radar was also 60 MHz and 8 kW but was incoherent (logarithmic detectors) with 250 ns or 60 ns pulse durations.

Examples: Targets of UTIG research in Antarctica that are relevant to Mars have included subglacial volcanics, subglacial water, and valley glaciers. The internal layering present in the Antarctic ice sheets is likely to be a good radar analog to the PLDs of Mars.

Enhanced geothermal flux. Evidence for active volcanism beneath the Antarctic ice has been previously described based on a combination of surface elevation, gravity, magnetics and radar sounding [6]. We have located what appears to be recently active volcanics based solely on radar sounding data (Figure 1) [8]. Evidence includes downwarped internal layers (from basal melting) correlated with a local topographic high and strong, flat radar echoes consistent with the presence of water draining into local topographic lows. Layer distortion from flow is ruled out since the feature is located near the Ross-Amundsen ice divide and the downwarping is contrary to layers expected from flow over a topographic high.

Subglacial Lakes. Subglacial lakes are known to exist in Antarctica, particularly beneath the thick, cold East Antarctic ice sheet [9]. Many dozens of small lakes of variable depth, stability and isolation exist there [10], and some of these are likely to be more analogous to potential subglacial features on Mars than the larger subglacial water bodies of Antarctica such as Lake Vostok. Figure 2 shows an example of such a lake, just a few kilometers across. Identification of such lakes has been automated by [10] based on several criteria obtainable directly from the radar data.

Valley Glaciers. Analogs to features that appear to be valley glaciers on Mars are found within the cold-based glaciers of the Dry Valleys [11, 12]. These

Figure 1. Evidence for locally elevated geothermal flux under the Antarctic Ice Sheet from UTIG 60 MHz radar sounding. The key indicator is the syncline in the layers whose amplitude correlates with depth, yet the axis is uncorrelated with bed topography. Bright basal echoes surrounding the target are consistent with water greater than 10 cm thick.
glaciers, whether on Earth or Mars, represent a substantial challenge for orbital radars due to their small size and the high surface clutter environments in which they are found. UTIG has acquired airborne radar sounding data over Taylor and Beacon Valleys in Antarctica and has developed techniques for mitigating the effects of surface clutter using those data [13].

Taylor Glacier flows from the East Antarctic ice plateau across the Transantarctic Mountains into Taylor Valley where it ends at ice-covered Lake Bonney. It has been considered to be frozen to the bed based on modeling of flow (no evidence for basal sliding). Following our efforts to identify the true subsurface reflectors near the terminus of the glacier, we applied echo strength analysis including a thermal model to determine dielectric losses [14]. We found that echo strengths at the base of Taylor Glacier are indeed not consistent with the presence of water.

In Beacon Valley, rock glaciers exist that appear nearly identical to viscous flow features observed on Mars [11, 15]. For reasons that are not yet clear, our radar sounding efforts over the rock glaciers of Beacon Valley have not penetrated through the rough surface.

**Conclusions:** UTIG’s legacy of airborne radar sounding in Antarctica will surely be useful to our efforts to understand Mars using similar “viewing-glasses.” If localized geothermal flux anomalies exist, or have previously existed beneath the PLDs, they may be detectable through the analysis of layer geometry in conjunction with careful consideration of possible ice dynamic effects. Analysis of echo character in conjunction with thermal models should enable the identification of subglacial liquid water if it exists there, and surface clutter mitigation will be a critical aspect in locations with significant surface topography.

**References:**

THE NORTH POLAR LAYERED DEPOSITS, MARS: TOPOGRAPHY, FLOW AND IMPLICATIONS FOR TIMESCALES.

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Introduction: The north polar layered deposits (PLD) of Mars consists mainly of H\textsubscript{2}O ice together with unknown amounts of dust and CO\textsubscript{2} ice [1]. It has formed over millions of years by deposition of layers of ice and dust. The current surface morphology is the result of climatic and geologic processes working over time, and the extensive layering visible in deep troughs suggest variations over time in the rates of dust and ice deposition.

The current large scale surface topography appears to be controlled mainly by mass balance [2, 3]. Yet the total mass balance of the north PLD is unknown, as is the role of flow in shaping the cap and thinning the layers.

In this work, we investigate the balance between mass balance processes and ice flow in forming the north PLD. We use the current surface morphology to examine the role of flow and to constrain the current mass balance. We discuss scenarios for north PLD evolution and the implications for the timescale of the north PLD layers.

Ice flow mechanisms: H\textsubscript{2}O ice is a polycrystalline material that flows when it is subject to stress. On Earth, ice flow is due to dislocation creep [4], while on Mars, it has been suggested that ice flow is due to grain boundary sliding [5]. Both mechanisms would allow ice to flow under very low stresses. For dislocation creep, in particular, the flow law is highly non-linear and the flow is very slow for low stresses. Therefore, ice is sometimes assumed to be perfectly plastic, i.e. there is no flow for driving stresses lower than a threshold value. For grain boundary sliding, however, the power of the flow law is 1.4, which is close to Newtonian flow with constant viscosity. The H\textsubscript{2}O ice in the north PLD must flow under gravity like terrestrial ice sheets, but the flow rate is slow because ice temperatures are colder, gravity is about a third of that on Earth, and flow may be governed by another mechanism.

Growth or shrinkage of non-equilibrium ice sheets: The terrestrial ice sheets in Greenland and Antarctica are relatively close to equilibrium. They have a high central area with accumulation of ice, and ablation in the lower marginal areas. Many smaller terrestrial ice caps and glaciers, however, are far from equilibrium, and this may also be the case for the Martian PLDs.

Weertman [6] has considered the rate of growth of non-equilibrium ice sheets far from the stable equilibrium state. According to his analysis, a growing ice sheet, which is building up with accumulation everywhere would eventually become large enough to flow, and would then adjust to the shape of a flowing ice sheet. A shrinking ice sheet which contain only one ablation zone covering the whole ice sheet, would become stagnant.

For an equilibrium ice sheet, the topographic shape is mainly controlled by the rheological properties and the flow law. Mass balance has a minor effect. For non-equilibrium ice sheets, however, the mass balance, and the mass balance history plays an important role.

Large scale topography of the north PLD: The surface topography of the north PLD was measured with high accuracy with the Mars Orbiter Laser Altimeter (MOLA) [7]. Figure 1 shows immediately that the topographic profile of the north PLD is not the profile of a flowing ice sheet. Indeed, Ivanov and Muhleman [2] found that the current large scale surface topography of the north PLD can be approximated by a sublimation model alone. Greve and Mahajan [3] showed that large scale ice flow rates are not significant compared to mass balance rates. A study of the upper 500 m of the north PLD stratigraphy reached a similar conclusion [8]. All these studies were done on the main part of the north PLD. A study of the profile along the west-facing tongue of Titania Lobe concluded, however, that the present topography (between troughs) is consistent with that of a flowing ice sheet [9].

Sublimation rates: Surface topography data shows, that a pattern of deep troughs are organized in a spiraling pattern around the pole. The spiraling polar troughs are thought to be formed in a combination of
sublimation along the south-facing slopes and wind effects. Ice flow is thought to be enhanced near the troughs because the surface slope is high, and the effect of ice flow is to smooth away the troughs.

We have used an ice flow model to calculate how much sublimation is needed to keep the troughs open [10]. Between troughs we have made different runs assuming little or no deposition. Our method provides a minimum constraint on the sublimation rates, but if the ice is stiffer than we have assumed, the minimum constraint would be smaller. We have integrated our estimated sublimation rates over the north PLD, and obtain a total sublimation in close accordance with estimates from atmospheric water vapor data [11]. Our results suggest further that troughs are forming in the current climate, that the cap is currently shrinking, and that sublimation rates are in the order of several cm/year in the troughs. In the deepest troughs, we obtain a sublimation rate constraint of 5 cm/yr.

**Discussion:**

*Formation of north PLD and the polar troughs:* A major shift in obliquity 5 Myr ago from a level around 35° to a level around 25° changed the solar insolation at the north pole [12]. It has been proposed that the north PLD could not have formed before 5 Myr, because the summer insolation was too high [13]. If the north PLD have built up continuously over the past 5 Myr, the average deposition rate at the north pole is given by the cap thickness/5 Myr. Assuming that the cap thickness is given by the MOLA profile, the average deposition rate is c. 0.5 mm/year, which agrees with estimates of deposition rate from stratigraphic analysis [12].

If the polar troughs have formed with the rates estimated from flow model constraints, deep troughs of 1km depth may have formed in the order of 10^5 years [10]. Further inspection of the solar insolation variation at the north pole [12], shows that the insolation has oscillated around the present level with amplitudes of about 50% of the mean value, but in the last 2-3·10^5 yrs the insolation has been relatively constant. Did troughs form during this period?

Troughs form along south-facing slopes of the north PLD. Are sublimation rates lower along the west-facing slopes of Titania Lobe, allowing flow to have a relatively more prominent role in shaping the cap there?

**Implications for timescales:** We have constructed a model of the north PLD. The model starts 5 Myr ago, and assume build up in first 4.75·10^6 yrs, then followed by trough formation in the last 0.25·10^6 Myrs. During the build up phase, the north PLD only gradually starts to flow. During the trough formation phase, sublimation rates varies greatly along south-facing slopes, and troughs grow and become separated by almost horizontal terraces. Flow may become increasingly more important near the troughs as the slope increases, but insignificant elsewhere. We show how layers are thinned by flow near troughs and elsewhere, and compare with other models of the north PLD [14].

**References:**

THE NORTH POLAR LAYERED DEPOSITS, MARS: SUBLIMATION RATES AND RECENT EVOLUTION. C. S. Hvidberg\textsuperscript{1}, N. B. Karlsson\textsuperscript{1}, and D. Tyler, Jr.\textsuperscript{2}\textsuperscript{1}Niels Bohr Institute, University of Copenhagen, Juliane Maries Vej 30, DK-2100 Copenhagen, Denmark, Email: ch@gyf.ku.dk, nannabk@fys.ku.dk
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Introduction: The north polar layered deposits (PLD) of Mars has formed by deposition of layers of ice and dust over millions of years. The current surface morphology is a combination of climatic and geologic processes working over time. The north PLD consists mainly of H\textsubscript{2}O ice together with dust and possible layers of CO\textsubscript{2} ice [1]. The north PLD is the most significant source and sink of atmospheric water vapor, and it interacts with the atmosphere through sublimation and deposition of H\textsubscript{2}O ice [2].

In this work, we investigate the effect of modeled surface winds on the current surface mass balance. We compare the modeled winds [3] with surface morphology of the north PLD, and we derive a sublimation map that include effects of surface winds. We discuss implications on the recent evolution of the north PLD and formation of polar troughs.

The large scale shape of the north PLD: The large scale shape of the main north PLD appears to be controlled mainly by sublimation. Ivanov and Muhleman [4] found that the current large scale surface topography of the north PLD can be approximated by a sublimation model alone. Several other studies have suggested that ice flow have had a minor role in forming the current large scale topography of the main north PLD [5,6]. For the south, Byrne and Ivanov concluded that the shape of the layer which they trace across 150 km of the south PLD is not consistent with typical shapes produced by flow [7].

Spiraling polar troughs: A characteristic feature of the north PLD, is the polar troughs that form a spiraling pattern centered around the north pole. The south facing walls of the troughs are steep and appear dark, while the horizontal surfaces between the troughs appear white. It has been suggested that deposition occurs at the white horizontal surfaces, while sublimation is enhanced at the dark south-facing slopes [8]. This alternating pattern of ablation and accumulation has later been termed accublation [9]. The alternating dark and white surfaces associated with the spiraling troughs suggest that there is a currently active interaction between the north PLD and the atmosphere associated with the troughs and involving deposition and sublimation of H\textsubscript{2}O ice.

Estimates of closure rates of the polar troughs by flow suggest a closing time of \(-10^5\text{-}10^6\) years, meaning that the troughs are relatively young features that have been forming during the recent and current climate [10,11].

The formation of the polar troughs is not fully understood. It has been suggested that the troughs are forming perpendicular to the catabatic wind field [12]. Indeed, Ng and Zuber have suggested how troughs could form as a result of an instability between albedo and wind feedbacks [13]. The spiraling structure was then proposed to arise from the direction of the catabatic wind flow, which is deflected by the Coriolis force due to the rotation of Mars and thereby forming a spiraling structure around the pole. Alternatively, it was suggested that a combination of accublation and ice flow could form spiraling troughs [14]. It has also been suggested that ice flow could play a role in controlling the depth of troughs [11].

Results: analysis of the surface morphology: The surface topography of the north PLD was measured with high accuracy with the Mars Orbiter Laser Altimeter (MOLA) [15]. We have analyzed the topography to obtain information about the large scale shape of the cap, and the characteristics of the spiraling polar troughs, including the orientation and depth of troughs, and the spacing between troughs. We find that steep slopes are consistently oriented towards SSW in agreement with previous investigations [12].

Results: Surface mass balance and surface winds: We have used a model to calculate the sublimation of water ice from the north PLD surface. We set up a surface energy balance, and assume that the total sublimation is a combination of buoyant diffusion of water away from a saturated near-surface layer, and turbulent mixing driven by local winds [16]. We use modeled local winds derived from a mesoscale model study of summertime atmospheric circulations in the north polar region [3].

Discussion: We discuss the correlation between the current large scale surface topography of the north PLD and the derived sublimation rates. We use an ice flow model to examine how the large scale topography correlate with the modeled summer surface wind field and the derived sublimation rates. We discuss the implication on the recent evolution of the north PLD.

We discuss the characteristics of the troughs and compare with the modeled summertime surface winds.
and derived sublimation rates. We discuss what the implications are on the trough formation process.

OBSERVATIONS OF THE LAYERING STRUCTURE IN THE SOUTH POLAR LAYERED DEPOSITS WITH THE MARSIS INSTRUMENT. A. B. Ivanov\textsuperscript{1}, A. Safaeinili\textsuperscript{1}, J. J. Plaut\textsuperscript{1}, S. M. Milkovich\textsuperscript{1}, G. Picardi\textsuperscript{2}, \textsuperscript{1}Jet Propulsion Laboratory, MS168-416, Pasadena, CA, 91106; e-mail : anton.ivanov@jpl.nasa.gov, \textsuperscript{2}Infocom Department, “La Sapienza” University of Rome, 00184, Rome, Italy.

Introduction

One of the many questions of Martian exploration is to uncover the history of Mars through analysis of the polar layered deposits (PLD) (extensive reviews in [1] and [2]). Martian polar ice caps contain most of the exposed water ice on the surface on Mars and yet their history and physical processes involved in their formation are unclear. This work will concentrate on analysis of the internal structure of the South Polar Layered Deposits using Mars Advanced Radar for Subsurface and Ionosphere Sounding (MARSIS) [3] experiment data.

Considerable advances have been made in recent years to understand the internal structure of the polar layered deposits. Researchers have employed large amount of imaging data returned by the Mars Global Surveyor and Mars Odyssey spacecraft. [4] have tested a hypothesis on the internal structure of the South Polar Layered Deposits (SPLD) and concluded that the top most layers are most likely parabolic in shape and therefore ablation and sublimation are most prominent forces in forming those layers. [5] have analyzed MOC images taken in the North Polar Layered Deposits (NPLD) and found that albedo patterns are consistent with 30m thick alternating units and that layer sequences are broadly distributed. [6] describes in detail possible mechanisms for formation of the basal unit, discovered by [7]. [8] expands this work, presenting an updated sedimentary history of the NPLD based on data from MOC and THEMIS instruments. Work in [8] clearly distinguishes between brighter units sitting unconformably on top of the dark layers. To explain observations in the SPLD, [9] have suggested an eolian-based model for accumulation and erosion in the Promethei Lingula region.

There is an ongoing debate on the extent the flow in the ice caps. Work by [4] in the SPLD has suggested that structure of the top portion of SPLD is not consistent with flow. Meanwhile work in [10], [11], [12], among others argue that part of the NPLD may have flown in the past. All investigations, including cited above, suggest a certain structure that would be observed according to presented models. The ultimate goal of work described in this abstract is to validate those hypotheses using MARSIS data.

The MARSIS instrument

MARSIS is a multi-frequency pulse-limited radar sounder [3], which uses synthetic aperture techniques to maximize signal-to-noise in the received data. MARSIS can be effectively operated at any altitude lower than 800 km in subsurface sounding mode, and below 1200 km in ionosphere sounding mode. The instrument consists of a dipole and a monopole antenna assemblies and an electronics assembly. Maximum penetration depths are achieved at the lowest frequencies. Penetration will be in the order of a few kilometers, depending on the nature of the material being sounded. On the dayside of Mars, the solar wind-induced ionosphere does not allow subsurface sounding at frequencies below approximately 3.5 MHz. In the subsurface sounding mode, MARSIS can operate at 4 different frequencies: 1.8, 3.0, 4.0 and 5.0 MHz, hence allowing a capability to operate in the presence of the ionosphere.

Data and Observations

The MARSIS instrument has already shed some light on the subsurface interface in the NPLD [3]. Now MARSIS provided very dense coverage of the subsurface interface and layering structure in the SPLD [13]. These observations are the first direct measurement of the subsurface structure of the Polar Layered Deposits.

Data from the MARSIS instrument can come in two basic types: standard data, which is processed on the spacecraft to reduce data rate and raw data, which contains raw echoes as received by the MARSIS instrument. Raw data mode, also known as, “individual echoes” mode, requires a much higher
REFERENCES

data volume and therefore is not used frequently. MARSIS allows simultaneous collection of standard data and individual echoes.

Figure 1 illustrates first results obtained by the MARSIS investigation [13] in SPLD. This figure illustrates the advantages of the “individual echoes” mode. Ground processing allows much greater resolution (30 m) along the track compared to 5 km for standard data products. The individual layers and the bottom interface are clearly seen in Figure 1. Note that if an internal layer is not parallel to the MEX tangential velocity over 5 km, it will be smeared and this what we observe with MARSIS in the standard data products (Figure 1, top panel). This “raw data” mode represents our best possibility at identifying thick (relative to an individual layer) sequences in the SPLD structure.

Summary
We will present MARSIS observations of the SPLD internal structure. Basis of our analysis will be MARSIS data taken in the “individual echoes” mode. We will attempt at correlating observed sequences with the layer exposures evident in the imaging data. This work will help to test hypothesis described in [4] on the internal structure of SPLD. Additionally, we will pay special attention to the Promethei Lingula region to compare MARSIS data with work described in [9].

References


INTERANNUAL VARIABILITY IN RESIDUAL SOUTH CAP ALBEDO: MOC OBSERVATIONS.  P. B. James1, B. P. Bonev2, and M. J. Wolff3, 1Space Science Institute, 4750 Walnut Street, Suite 205 Boulder, CO 80301; mailto:pjames@spacescience.org, 2Goddard Space Flight Center, Greenbelt, MD 20771 bbonev@ssedmail.gsfc.nasa.gov, 3 Space Science Institute, 4750 Walnut Street, Suite 205 Boulder, CO 80301; mailto:wolff@spacescience.org.

MGS has now observed four complete spring-summer recessions of the Martian south polar cap. High-resolution MOC images of the residual south polar cap of Mars (RSPC) acquired during the first year revealed a surface that was intricately sculpted into multitude of unique surface features, often called “Swiss cheese,” etched into about six identifiable layers of material [1]. Images acquired during the second year showed that many of these pits had expanded during the intervening year [2], indicating current erosion of residual CO2 frost. Detailed study of observations for the first three years revealed that the RSPC actually consists of several identifiable units and that some of these units show evidence of deposition and erosion on the scale of ~ 100 years [3].

One of the mechanisms that could trigger net erosion or net accumulation during a year in the RSPC is the response of the cap to changes in the radiation flux absorbed by the cap. Such changes could be caused by interannual changes in the seasonally dependent surface albedo of the cap, e.g. due to variable deposition of dust and water, or to changes in the dust content of the atmosphere that redistribute the energy incident on the cap. MOC wide-angle cameras imaged the RSPC at certain seasonal (Ls) dates in each of the four years. The visual appearances of the residual cap in the four years observed by MGS show almost no variation. Intercomparisons of the fixed Ls MOC views of the RSPC are examined here to investigate the question of whether the albedo of the RSPC is constant or shows the effects of the small scale erosion.

The behavior of the albedo of the seasonal ice deposits in the south polar cap is complicated. Several studies, starting with that of Page [4], indicate that the frost deposits actually brighten in proportion to insolation in the spring and reach a maximum slightly after summer solstice. The observed high albedo of the seasonal cap during the periods of maximum insolation [5] is the root cause of the somewhat countintuitive preservation of perennial frost in the south. In the central portion of the cap, i.e. the RSPC, there is a drop in albedo in mid-summer that possibly reflects the removal of the seasonal CO2 frost deposits, exposing the residual cap terrain. The albedo of the actual residual deposits is less than that of the seasonal deposits that are deposited over them.

Figure 1 shows the albedos extracted from wide-angle red filter MOC images (600nm) of the RSPC region (both bare ground and condensate) acquired at Ls = 340° for the four years observed by MOC, i.e. Mars years (MY) 24-27 (year 1 begins on April 11,1955) [6]. It can be seen in the figure that there is considerable variation in the albedo of the RSPC from one Mars year to another, but that the variation is not monotonic. The lowest albedo is found in MY26 (the third year of MGS mapping), but the albedo in the fourth year of mapping is significantly larger. Thus the low albedo in the third year does not seem to have had any obvious feedback effect on the following year. This contradicts the hypothesis that the RSPC albedo is correlated with the physical changes in the “Swiss cheese” features in the cap that appeared to erode monotonically during the first three mapping years as mentioned above. The highest albedo is clearly in MY25, the year of the large 2001 dust storm that started near equinox.

Figure 1: Histogram of Lambert albedo for the RSPC region for four Mars years: MY24 (blue), MY25 (green), MY26 (orange), and MY27 (violet). The larger peak represents unfrosted ground in the images, while the smaller peaks are residual CO2 frost. There is no secular increase or decrease of albedo from year to year; the second year has the highest cap albedo, and the third year has the lowest. The abscissa scale is 255 x the Lambert albedo.

We have considered three hypotheses of processes that could produce large changes in RSPC albedo that are not correlated with the observed cap erosion and that are amenable to testing:

1. The 2001 dust storm had little effect on the large scale cap regression [7], and any radia-
tive effects of the dust on the high albedo RSPC region would have hastened the sublimation of the seasonal CO2 deposits [8]. Perhaps the dust had effects through other mechanisms such as providing nuclei for enhanced condensation near equinox. Even small variability in the seasonal CO2 cycle in the core of the South Polar Cap could result in fairly large variations in the date that the residual CO2 frost is uncovered because of the low rate of net sublimation at that time. In this scenario, the date at which the seasonal CO2 disappears should correlate with the large scale, average albedo of the cap.

2. Atmospheric dust removes incident solar flux and affects the measured albedo of the RSPC. In other words, variability in the opacity of dust over the RSPC region in late summer could bring about the observed interannual and local variability in the apparent albedo. This scenario can be tested using radiative transfer models with reasonable variations in opacity.

3. Emission angles of a fixed location in the RSPC can vary substantially depending on the viewing geometry. The nadir point is generally near -87° latitude when MGS images the RSPC; because the cap is displaced from the geographic pole, a feature located at -87° could be as much as ~ 6° from nadir where the emission angle would be ~60°. So a departure of the surface phase function of the residual ice from the Lambertian form assumed here would then show up in the Lambert albedo. If only this effect were involved, albedo variations should vanish at the geographic pole; the persistence of the albedo differences for a small region centered on the pole indicates that the changes in measured albedo are not due to phase function alone.

In summary, the Lambert albedo of the RSPC varies substantially from one Martian summer to the next, and the variability is not obviously correlated with the smaller scale erosion of the cap surface. It appears possible that the albedo changes may be related to subtle variations in the seasonal CO2 cover from one year to the next.

Acknowledgement: This work was supported by a Mars Data Analysis Grant

DISPLACEMENT OF MARTIAN SURFACE DUE TO SEASONAL SURFACE MASS REDISTRIBUTION AND ITS DETECTION FROM LANDER-ORBITER-EARTH RADIO LINKS., Ö. Karatekin¹, J. Hagedoorn¹, ², T. Van Hoolst and V. Dehant, ¹Royal Observatory of Belgium (3, Av. Circulaire 1180 Brussels Belgium. o.karatekin@oma.be), ²GeoForschungsZentrum Potsdam (Telegrafenberg A17, D-14473 Potsdam, Germany (jan@gfz-postdam.de)

Introduction: The CO₂ atmosphere of Mars sublimes and condenses on seasonal time scales, resulting in a large scale surface mass re-distribution. The surface of the planet deforms due to this periodic surface loading. In the present study, we calculate radial surface displacements by using a visco-elastic approach for several interior models of Mars and investigate the possibility of the detection of this periodic displacement with lander-orbiter-Earth radio links.

Calculations: The visco-elastic deformation of Mars caused by the time-variable load is simulated by computing the resulting perturbations for a spherical, self-gravitating, incompressible, Maxwell visco-elastic continuum. The corresponding boundary value problem (BVP) is solved by the spectral finite-element representation developed by [1]. In this representation, the equations are stated in the integral formulation, with the angular dependence represented by spherical harmonics and the radial dependence by finite elements. The characteristic feature here is that the BVP is solved in the time domain.

For our calculations we describe Mars as a liquid core, surrounded by a visco-elastic mantle overlaid by an elastic lithosphere. The elastic parameterization is given by radial dependent profiles of density, and shear and bulk modulus. The viscosity of the mantle is considered as a variable of the problem (10²³>µ>10²⁰ Pa.s.). The surface mass density variations are taken from LMD GCM database 4.1 [2].

Results and Implications for Future Measurements: The largest surface displacements occur at high latitudes where surface CO₂ density varies in seasonal time-scales. This is in agreement with the expected degree-one variations of the gravitation field of Mars [3][4]. The peak to peak semi-annual amplitude of the radial displacements depends on the internal structure of the planet (mainly to the viscosity of the upper mantle) and be over 100 mm in the vicinity of the Polar Regions.

The resulting radial displacements due to surface mass variations are about one order of magnitude larger than to the tides caused by Sun [5]. These displacements can be easily detected from a precise (two way X and/or Ka band) lander-orbiter-Earth radio links. Provided that the polar lander is equipped with such a transponder, the analysis of the measurements could be used to constrain the viscosity of the upper mantle if the surface loading (seasonal CO₂ cycle) is sufficiently well known.

Moreover, precise lander-orbiter-Earth radio links can improve the orientation parameters of Mars, which are directly related to the interior structure. The knowledge of the present state and size of the core and the mantle structure is crucial for the understanding of the evolution of Mars. Another significant outcome of such measurements would be that the better precession rate which would improve significantly the obliquity evolution of Mars over tens of millions of years [6].

MASS AND DENSITY OF SEASONAL POLAR DEPOSITS FROM TIME-VARIABLE GRAVITY DATA.
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Introduction: There are several possibilities to estimate the global scale mass exchange on Mars [1][2]. Detection of temporal changes in the long wavelengths of the Martian gravitational field is one of the most direct methods as it does not need physical models related to the surface/atmosphere interactions or to the subsurface modeling. It also offers the advantage of continuous data for many years as tracking data will be available for any Mars orbiter.

The mass exchanged between the atmosphere and polar caps can be estimated from the temporal variations of the zonal gravity coefficients of degree two and three, $\Delta C_{20}$ and $\Delta C_{30}$ [3][4][5]. We determined the mass and pressure variations on seasonal time scales from the reported orbital analysis of Mars Global Surveyor (MGS) and Mars Odyssey spacecraft data [6][7]. The results show good agreement with High Energy Neutron Detector (HEND) observations and the numerical models of the atmosphere dynamics of NASA Ames and LMD [5]. In the present study, we present the mass variations in north and south polar caps deduced from time-variable gravity field. In addition, we combine the mass variations with the MOLA altimeter data [6] to obtain the densities of the seasonal polar deposits.

Method: The mass exchange between the surface and atmosphere causes changes in the zonal Stokes coefficients $C_{n0}$ (of degree $n$ and order 0) of the gravity field. The orbit of the spacecraft around Mars is sensitive mostly to the variations of the long wavelength components of the gravity field which are associated with the global-scale seasonal mass exchange. The variations in the spacecraft orbit, calculated from MGS and Mars Odyssey tracking data analysis, yield the time-variable gravity coefficients of $\Delta C_{20}$ and $\Delta C_{30}$ [6][7][8][9].

The zonal coefficients $\Delta C_{20}$ and $\Delta C_{30}$ can be expressed as an integral over the seasonal surface mass density variations, $\Delta \sigma$. It can be shown that the mass variations on the north and south hemispheres ($\Delta M_N$ and $\Delta M_S$), are linearly related to $\Delta C_{20}$ and $\Delta C_{30}$ [5]. Three approximations are made: (1) the elastic yielding of the planet under the surface loading and any seasonal mass variation inside the planet are neglected. The effect of elastic yielding is less than 1% [10]. (2) $\Delta \sigma$ is assumed to be a function of colatitude only. This is a very reasonable assumption since, during the cap retreats and advances, the caps remains mostly axisymmetric, except when the surfaces of seasonal deposits are close to minimum. (3) atmospheric pressure contributions to $\Delta \sigma$ are neglected. On the basis of GCM simulations of the Martian atmosphere, this would introduce errors less than 6% and 2% for $\Delta C_{20}$ and $\Delta C_{30}$, respectively [11].

Mass variations: The mass variations in the north and south polar caps, $\Delta M_N$ and $\Delta M_S$ for linear $\Delta \sigma$ distributions can be expressed in the form of annual and semi-annual periodic series [5].

$$\Delta M_N = 1.353 + 1.895 \sin (L_s + 109.3) + 0.615 \sin (2 L_s + 120.4)$$
$$\Delta M_S = 2.446 + 2.761 \sin (L_s - 59.1) + 0.615 \sin (2 L_s + 120.4),$$

where $L_s$ is the solar longitude. The coefficients in the above equations are in units of $10^{15}$ kg and the phases are in degrees. We used only the MGS data [4][5] and introduced the variations of the polar cap boundaries from the outputs of the NASA Ames GCM model. The mass variations are shown in Figure 1. The reported formal errors for $\Delta C_{20}$ and $\Delta C_{30}$ solutions yield $\Delta M \approx \pm 0.34 \times 10^{15}$ kg. The solution can be improved by adding the recent data from MGS and Mars Odyssey [8][9].

Density variations: There are several independent estimations of the density of the Martian polar caps. The results reveal several interesting deviations. The combined Mars Orbiter Laser Altimeter (MOLA) and MGS gravity data ($\Delta C_{20}$) give a mean cap density of $910 \pm 230$ kg/m$^3$ for both poles [6]. MGS Thermal Emission Spectrometer (TES) assumes a mean density

![Figure 1: Seasonal mass variations in the North and South poles.](image-url)
of 1606 kg/m$^3$ as an upper limit. Estimates of Martian northern CO$_2$ ice and snow/ice mixture densities based on combining TES mass balance calculations and MOLA seasonal elevation changes are 1107±150 kg/m$^3$ and 979±133 kg/m$^3$, respectively [12]. MOLA elevations changes combined with the Gamma Ray Spectrometer and Neutron Spectrometer data (GRS) for the northern cap yield much lower densities (∼500-600 kg/m$^3$) [13][14]. The solution of GRS compared to the previous MOLA and TES estimates are favored [2] partly (among other reasons) due to their good agreement with the NASA Ames GCM results [13]. The combined analysis of HEND and MOLA data, on the other hand, yield densities in the range 900-1100 kg/m$^3$ [15].

We calculate the density variations by combining our solutions of $\Delta M_S$ and $\Delta M_N$ with the MOLA data. The variations of the cap volumes are taken from the MOLA elevation variations given in [6]. The elevation data is more consistent with the assumed linear variation of $\Delta \sigma$ than the constant $\Delta \sigma$ and point mass models discussed in [5] for the mass estimation. We consider only the elevation changes from ±60° poleward. The precision of the MOLA elevations are reported as ±0.5m and ±0.6m for the north and south caps [6]. Our analysis reveals densities of 1124 ± 690 kg/m$^3$ and 1214 ± 258 kg/m$^3$ for the south and north caps respectively. Data points with large relative errors are neglected, restricting the current analysis to the periods when the ice cap elevations are near maximum. The uncertainties of MOLA depth estimates are larger in the south pole due to possible systematic errors [12]. Most of the previous density estimates did not consider the south pole [12][13][14][15]. A detailed analysis of the uncertainties of MOLA elevation for the north pole is given in [14]. The polar cap densities of the present study are larger than those deduced from the combined GRS/MOLA data analysis [13][14] but in good agreement with the others [6][12][15].


Figure 2. Density estimates for the south and north polar caps from mass variations deduced from time-variable gravity data and the volume changes calculated from MOLA elevation data.
Ice caps on Mars and on Kilimanjaro: Potential for Analogies

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The north polar ice cap on Mars is separated into a system of horizontal terraces and steep southward facing scarps with a surface slope of 10-15°. The scarps are dark, while the horizontal areas are white. Pictures show that the dark scarps have an extensive layering, which can be followed over hundreds of kilometers. The scarps form a spiral pattern around the pole. It has been suggested that ice sublimates along the dark scarps, because they receive most solar insolation, and the layers are visible, while the white terraces accumulate ice. The ice caps are thought to be millions of years old, and the stratigraphy is thought to show past climate changes on Mars.

The ice remnants sitting on the summit plateau of the high standing, tropical, dry and cold Kilimanjaro in East Africa can be characterized in a similar way, though they are very different in size. Due to the equatorial setting, the pronounced cliffs that margin the Kilimanjaro tabular summit ice do not only face south such on the Mars' pole, but south and north, parallel to the Earth's equator. Solar radiation drives their constant and irreversible retreat, sublimation and small precipitation amounts govern the mostly balanced mass budget on the horizontal terraces. Since 2005 we maintain a detailed setting of energy and mass balance instruments on Kilimanjaro and measure fluxes from and to surfaces of different inclination. This has taken us to a thorough understanding of how different processes maintain the observed features and the understanding allows an improved interpretation of the history of glaciers and the related climate drivers on Kilimanjaro.

We offer our understanding and findings for a potentially advanced interpretation of the behaviour and the history of Mars' north polar ice caps.
**The Robotic Arm Camera (RAC)** H.U. Keller¹, W.J. Markiewicz, S. Hviid, M.T. Lemmon², P. H. Smith³, R.V. Morris⁴ and the Phoenix Science Team,¹ Max-Planck-Institut für Sonnensystemforschung, 37191 Katlenburg-Lindau, Germany, email:keller@mps.mpg.de, ²Department of Atmospheric Science, Texas A & M University, College Station, TX 77843-3150, USA, and ³Department of Planetary Sciences, University of Arizona, Tucson AZ 85721, USA, ⁴NASA Jhonson Space Center, Houston, TX, 77058, USA.

**Introduction**: The Robotic Arm Camera (RAC) is one of the key instruments of the Phoenix lander required for science analysis and operations. This lightweight instrument employs a front lens with variable focus range and takes images at distances from 11 mm (image scale 1:1) to infinity. Colour images with a resolution of better than 50 µm will be obtained to characterize the Martian soil. Spectral information of nearby objects is retrieved through illumination with blue, green, and red lamp sets. The optical microscope (OM) within the MECA instrument set uses the same CCD detector and common readout electronics. This hardware is supplied by Max Planck Institute for Solar System Research. A detailed description of the RAC for the Mars Polar Lander is given by [1].

![Figure 1. RAC mounted on the robotic arm.](image1)

**Science objectives**: The RAC is mounted close to the scoop at the “wrist” of the robotic arm. Its task is to
- characterize the area of digging
- search for near surface ground ice in the trench while digging
- characterize the physical properties of the soil (grain size, clumpiness, heterogeneity) that is transported by the scoop to TEGA and MECA
- document the origin of the soil samples
- search for seasonal and climate records in the form of layering in the trenches
- observe the surface and subsurface condensation and sublimation of water

The RAC’s optical design implemented a variable focus. Close up images of material in the scoop can be taken down to a scale of 23 µm per pixel. The field of view of the RAC is 54° x 27° so that it could be used as a backup stereo imager by pointing the robotic arm. At infinity the scale is 1.8 mrad per pixel.

![Figure 2. RAC images of material in the RA scoop.](image2)

The soil picked up can be characterized inside the scoop. Changing the distance of best focus from the front to the back of the scoop allows to determine the contents in 3 dimensions by combining the images. In this way the volume of the soil can be estimated. The clumpiness of the soil and its grain size distribution can be documented. Images illuminated by the colour LED will allow us to distinct different mineralogic prominences. The careful documentation of the samples from their origin to their physical and mineralogical characteristics is of importance for the further analysis by MECA and TEGA. We will demonstrate the capabilities of the RAC using images taken during calibration and with earlier models.

Figure 3. Rock slap imaged at 23μm/pixel resolution using the flight unit RAC camera.

Figure 4. Stereo image of a trench (about 50 cm wide) made with RAC.
GEOLOGIC MAPPING RESULTS OF THE AUSTRALE SCOPELI REGION WITHIN PLANUM AUSTRALE, MARS. E.J. Kolb¹ and K.L. Tanaka¹,², Arizona State University, Dept. of Geological Sciences, Tempe, AZ 85287, eric.kolb@asu.edu, ¹²Astrogeology Team, U.S. Geological Survey, Flagstaff, AZ 86001, ktanaka@usgs.gov.

Introduction: Australe Scopuli forms a broad section of the Martian south polar ice plateau, Planum Australe, within an area southwest of Cavi Angusti. The region consists of deep canyons associated with the curvilinear canyon system that cuts Australe Mensa, the ice cap’s dome-shaped plateau centered at 87ºS, 356ºE. The canyons expose the south polar layered deposits (SPLD) that make up Planum Australe.

Our preliminary MGS and ODY-based regional geologic mapping [1] divides the SPLD into the Planum Australe 1 and 2 units. The Planum Australe 1 unit comprises most of the volume of the SPLD and is characterized as evenly bedded sequences of meters-thick layers exposed in canyon walls. The Planum Australe 2 unit is also comprised of multiple layers that are evenly-bedded although the unit’s total thickness is <300 m. Individual layers generally appear slightly thicker than their Planum Australe 1 counterpart and often exhibit moderately to heavily pitted and knobby surfaces. The unit unconformably buries the eroded topography of the Planum Australe 1 unit and is largely preserved on polar surfaces other than the canyon scars. Our mapping of the Promethei and Ullitum Chasmata complex [2] identifies a regional unconformity within the Planum Australe 1 unit allowing further subdivision of the unit into the 1a and 1b members. The orientation and outcrop expression of the unconformity demonstrates that the chasmata formed early in the SPLD accumulation history, likely by aeolian downcutting of surface depressions in the SPLD during erosional episodes. The mapping also demonstrates that the curvilinear canyons are stable features that have not migrated “conveyor-belt-style” along the plateau surface either by glacio-dynamic or ablation-driven processes [see ref. in 2].

Mapping bases used in this study include the MOLA 1/512° (115 m/pixel) topographic grid, southern spring and summer THEMIS visible-image mosaics (72 and 36 m/pixel, respectively; [3]) and southern summer MOC narrow-angle images (2-14 m/pixel; through orbit S04). Also, we generated a southern summer THEMIS VIS mosaic at full resolution (18 m/pixel) of the region using images from orbits 7955 through 9752. These data sets permit us to map, characterize, and evaluate the region’s geologic materials, features, mechanisms, and history.

Results and Discussion: The study area (Fig. 1) contains six curvilinear canyons and a high-standing plateau (informally labeled a) of SPLD cut both by several small canyons and elongated semi-closed depressions. The canyons and depressions trend northeast, extending for tens of to >250 km. SPLD that comprise the canyon scarps are well exposed. Within individual canyons, wall slope does not vary greatly over lateral distances, thus the outcrop expression of SPLD sequences remains consistent along an exposure. As a result, individual layers can be traced along the entire outcrop. The SPLD outcrops do not contain marker beds (e.g., exceptionally thick layers, distinct layer weathering patterns, layers with notably darker or brighter albedo) suitable for intra-canyon stratigraphic comparisons, which thus far has precluded inter-canyon correlation of layer stratigraphy. Subdivision of SPLD sequences is based on bedding unconformities and on their bench or terrace outcrop expression, analogous to the approach we used in mapping of SPLD within Australe Sulci [2]. We delineate Planum Australe 1 unit into members a and b (relationship to similar members in [2] is unknown), and also map the bases of sequences (labeled a₁ to a₅) defined by prominent layers within member a. Sequential numbering of the sequences corresponds to the highest elevation each sequence is found at within the map extent, Fig. 2).

The α canyon cuts to the base of Planum Australe, and its scars expose laterally continuous SPLD sequences devoid of obvious unconformities. Therefore, we assume that the SPLD exposed in α canyon is representative of the basal kilometer of SPLD that comprise Planum Australe. At the head of α canyon, the terrace-forming a₁ and a₅ sequences each crop out at the same elevation on each side of the canyon. Within the northeast extent of the δ canyon, the canyon floor exposes substrate, and the a₂ sequence is at the same elevation as the a₁ sequence within α (Fig 2). Therefore, we assume that the a₁ and a₂ sequences are equivalent.

Within α, δ, and β, unconformable bedding structures are not observed within the a₁₃ sequence exposures, indicating that deposition was not interrupted by erosion through a₃ sequence emplacement. Local unconformities are observed in the lowest sections of the κ and λ canyons intersecting the apices of substrate prominences. This geometry indicates the unconformities formed by planation of high-standing polar deposits that buried the prominences during apparent short periods of non-deposition.

Within the κ and λ canyons, the a₄ basal contact is unconformable along much of its exposure, cutting across a 200-m-thick stack of underlying SPLD and suggesting at least a comparable amount was removed from this region prior to a₄ emplacement. The spatial extent indicates erosion was not uniform, but preferentially removed SPLD from the eastern parts of Australe Scopuli and also equatorward.

Within ω canyon at the nexus of the λ and α features, deposition of sequence a₅ over an uneven paleosurface resulted in variably dipping beds within the a₅₅ sequences. Within the same canyon, member b beds unconformably bury the a₅₅ sequences. Similarly, member b unconformably buries sequence a₃ in the β canyon. These relationships indicate that canyon development began prior to member b deposition. At a₅, the stacked occurrence of unconformities within stratigraphically higher sequences indicates that the canyon has undergone down-cutting during each of at least three depositional hiatuses. This suggests a formational history similar to that of Promethei and Ullitum Chasmata, where substrate-controlled depressions (seedling chasmata) were progressively enlarged by funneling of katabatic winds during SPLD depositional hiatuses [2].

The uneven extent of member b indicates that intense erosion within eastern Australe Scopuli followed member b emplacement, removing most sequences of the Planum Australe 1 unit that are younger than a₄ from areas that comprise the κ and λ canyons and the northeast section of the ω canyon. For each erosional episode, wind action is the likely erosional agent. Finally, the Planum Australe 2 unit was emplaced and then was partly eroded to its current extent.

We identify 28 candidate small impact craters and nine buttes. Their mean diameter is 300 m, and all have diameters
~1 km. The buttes are morphologically similar to buttes within Promethei Lingula interpreted as degraded, perhaps exhumed, pedestal craters [2]. Their density indicates a relatively long exposure age for the member a and b surfaces.

**Summary:** The geologic history of the Australe Scopuli region is as follows: (1) continuous deposition of sequences a1 through a3, (2) preferential removal of a1-3 form eastern Australe Scopuli, (3) emplacement of a4 for which the eroded paleosurface controls the location of dipping beds and canyons within the α plateau, (4) regional deflation and likely initiation of canyons, (5) deposition of member b followed by regional deflation that was most intense within eastern Australe Scopuli, (6) long hiatus between deflation and Planum Australe unit 2 emplacement, and (7) subsequent erosion of all SPLD material until now.

**Future work:** We are extending this mapping west into canyon systems that connect with the Promethei and Ultimum Chasmata. When completed, we will determine the stratigraphic relationships and placement of SPLD units, members, and sequences within the chasmata.


Fig. 1. Geologic map of Australe Scopuli region of Planum Australe, Mars. Units include: Salmon and blue areas = members a and b, respectively, of the Planum Australe 1 unit; green = Planum Australe 2 unit; purple = substrate. Colored lines are traces of member a sequences discussed in text; blue = a1 (dotted where buried), green = a2, yellow = a3, orange = a4, red = a5; heavy dashed lines mark unconformable bedding structures as seen in THEMIS visible images. Red dots mark impact craters; green dots mark circular buttes that may be degraded pedestal craters. MOLA shaded relief base (115 m/pixel); 5º lat/long grid, south pole at upper right.

Fig. 2. Cross-section A-A'. 25X vertical exaggeration. See Fig. 1 for explanation of unit and line colors.
Conditions for water ice accumulation within craters on Mars. K.J. Kossacki\textsuperscript{1}, W.J. Markiewicz\textsuperscript{2}, G. Portyankina\textsuperscript{2}, G. Neukum\textsuperscript{3}, and HRSC team, \textsuperscript{1} Inst. of Geophysics of Warsaw University, Warsaw, Poland, kjkos-sac@igf.fuw.edu.pl, \textsuperscript{2}Max-Planck-Inst. for Solar System Research, Katlenburg-Lindau, Germany, \textsuperscript{3}Institute for Geo-sciences, FUB, Berlin

Introduction: Images of a Martian crater show intriguing distribution of water ice. The crater is at 71°N, 103°E. At the floor of this crater an almost circular region of water ice is observed while the rest of the crater floor appears defrosted. This effect was first found in the MOC image M23-01916 Wide Angle Camera taken during late Martian Spring, at solar longitude L_s = 110°. A year later, at L_s = 154°, an image taken by HRSC camera on Mars Express shows the same effect (Figure below). MOC images taken during seasons more near winter show relatively uniform coverage of CO\textsubscript{2} ice.

This type of preferential frosting in the middle of a crater has also been seen in HRSC images of other craters.

We are interested to quantify what conditions are needed so that the observed distribution of ice cover on the crater floor appears only seasonally in some locations and when is it stable on a longer time scale. We model seasonal cycle of condensation and sublimation of CO\textsubscript{2} and H\textsubscript{2}O ices, within the crater at the observed location. Our simulations show, that the formation of a long lasting, or permanent ice layer in the center of the crater is possible only when below it the regolith has thermal inertia higher than the rest of the crater. This is the case for many Martian craters where the central region is made up of old highly consolidated dune field, while the rest of the crater floor is covered by younger fine-grained deposits.

Model: The numerical model used in this work describes non-uniform seasonal condensation and sublimation cycle of ices within a Martian crater. We include in the model the time dependent illumination, infrared emission, heat transport within the regolith and the energy losses due to sublimation or condensation. The flux of absorbed light depends on the local orientation of the facet (wall or bottom) and on the current position of the Sun. The latter evolves due to orbital motion and rotation of Mars. We consider direct solar light, as well as light scattered once within the crater. Condensation or sublimation of H\textsubscript{2}O is determined by the local surface temperature and the current near-surface density of vapour in the atmosphere taken from the LMD/Oxford GCM results. Illumination of the depression is calculated in three dimensions. The heat diffusion equations are solved in cylindrical coordinates, but for computational reasons only in two dimensions: vertical and radial. These calculations were performed for eight directions from the center of the crater.

Results: We have found, that the accumulation of H\textsubscript{2}O ice on the crater bottom is only possible if the thermal inertia of the presumed dune below it is significantly larger than that of the most of the crater floor covered by fine dust. In Figure below we show profiles of the surface density of H\textsubscript{2}O ice in the center of the crater, and at the $\frac{3}{4}$ of the radius in the direction to the pole.
**Summary:** Different types of material covering the crater bottom, and hence non-uniform thermal inertia have significant impact on the local condensation/sublimation of ices. With realistic values of thermal inertia of fine Martian dust and highly consolidated dunes we can explain observed distribution of ice on the crater bottom. This is further demonstrated in the figure below which shows several snapshots of results from our model.
THE 2007 PHOENIX MARS SCOUT WET CHEMISTRY LABORATORY: STUDYING THE CHEMISTRY OF THE POLAR SOIL/ICE.  S. P. Kounaves¹, M. H. Hecht², & The Phoenix Team, ¹Tufts University, Department of Chemistry, Medford, MA 02155  Tel: 617-627-3124, ²Jet Propulsion Laboratory, 4800 Oak Grove Dr., Pasadena, CA 91109

Introduction. The vast northern polar lowlands of Mars, known as Vastitas Borealis, could contain a chronological record of what may have an ocean, its remnant evaporitic stratigraphy, deposits of volcanic eruptions, or the advances and retreats of the polar ice cap. Recently, the Mars Odyssey orbiter has shown the existence of large amounts of subsurface water ice in the northern latitudes, buried under several to a few tens of centimeters of dry regolith [1]. The interactions between atmosphere, water, dust, and subsurface ice, in conjunction with the periodic changes of the ice/soil boundary, have no doubt produced unique chemical signatures in the regolith.

The 2007 Phoenix Mars Scout lander is designed to touch down at the high northern latitudes of Mars, between 65-72°N. It will acquire and analyze samples of soil and ice, and monitor atmospheric conditions. Its major goal is to unravel the history of water in all its phases and the record left in the geochemistry of the regolith. In addition, the Phoenix will address bio-habitability by, identifying potential chemical energy sources available to support life, determining whether the subsurface geochemistry is hostile to life, and identifying the potential of the geochemical environment to preserve paleontological evidence.

The Wet Chemistry Lab. To help analyze and interpret the chemical record, the Phoenix carries with it a variety of analytical instruments [2,3]. Foremost in helping us to understand the chemical record will be four single-use independent wet chemistry labs. The Wet Chemistry Labs (WCL) [4], originally developed as part of the Mars Environmental Compatibility Assessment (MECA) package for the cancelled 2001 Mars Surveyor Program lander, were rebuilt for the 2007 Phoenix mission. Each WCL (shown below) consists of a lower “beaker” containing a set of chemical sensors (inset image) designed to analyze the chemical properties of the regolith, and an upper “actuator”, for adding the soil, reagents, and stirring.

The beaker assembly contains an array of sensors consisting of solid state and PVC (polyvinyl chloride) membrane based ion selective electrodes (ISE). These sensors will analyze for inorganic anions and cations, including calcium, sodium, potassium, magnesium, chloride, bromide, ammonium, nitrate, lithium, and sulfate. The array also includes special electrodes for pH, conductivity, oxidation-reduction potential (Eh), anodic stripping voltammetry (ASV) for heavy metals such as Cu²⁺, Cd²⁺, Pb²⁺, Hg²⁺, chronopotentiometry (CP) for independent determination of chloride, bromide & iodide, and cyclic voltammetry (CV) for identifying and analyzing possible reversible and irreversible redox couples.

The upper actuator assembly consists of a sealed Teflon-coated titanium water tank with a puncture valve, a sample loading drawer, a screened funnel, a stirrer motor with impeller, a pressure sensor, and a five-crucible reagent dispenser. The pressurized tank holds 25 mL of solution that serves to leach the soluble components from the soil and also contains the first calibration standards for the ion selective electrodes and, in three of the WLCs, the background electrolyte (LiNO₃) for the lithium reference electrodes. The first of the five crucibles contains a second addition of calibration standards, the second contains 2-nitrobenzoic acid, and other three contain barium nitrate for use in determining sulfate. The sample loading “drawer”, containing a compartment which holds approximately 1 cm³ of soil, is designed to receive the soil from the robotic arm, remove excess soil, and deposit it into the beaker. A funnel with a wire screen allows only particles < 2 mm to fall into the drawer.

The data provided by the WCLs from each sample analysis will provide a variety of chemical and physical parameters to help us understand the geochemistry of Mars’ polar environment and its history.

INTRODUCTION: The spatial pattern of accumulation over an ice cap is a fundamental parameter when inferring the ice-flow and climate histories. Internal layers, taken to be isochrones, contain information about accumulation patterns in both space and time. The large body of radar data from terrestrial ice caps has greatly increased our understanding of ice-sheet evolution and climate.

Deeper, older layers reflect conditions further in the past, but they have been more affected by horizontal gradients in strain rate and accumulation. The depth of a deep internal layer does not directly reflect the accumulation rate at that point. Therefore, this information is highly valuable but more difficult to interpret. Formally solving this inverse problem is necessary to determine the correct accumulation pattern recorded by deep internal layers.

Internal layers have recently been observed on the Martian ice caps by the MARSIS ice-penetrating radar [2], and we expect that more layers will be observed with SHARAD. Assuming there was a time when ice-flow controlled the shape of the Martian ice caps, we can recover some of the information which remains in the deep internal layers. Waddington et al. [1] solved this inverse problem on terrestrial ice sheets. Here, we apply the method to Martian ice.

METHOD: A steady-state flow-band model is used to calculate layer evolution [1]. Layer prediction in this forward calculation depends on the layer age, and on the accumulation-rate pattern. Inverse methods are used to find physically reasonable values of these unknown parameters. The preferred parameters generate an internal layer that fits the data within a defined tolerance.

Figure 1. Flow-band geometry showing surface elevation (S), bed elevation (B), ice thickness (H), and flow-band width (W).

FORWARD MODEL. Figure 1 shows a typical flow band. The forward model uses the governing equations for ice flow along with boundary conditions, and parameter values, to calculate the steady-state ice-surface topography and flow field everywhere in the flow band. Particle paths are found by integrating the velocity field. By following a number of particles over time, we can map out an internal layer of a particular age.

Figure 2. Antarctic radar profile along model flow band. The dashed line denotes the targeted internal layer (taken from [1]).

INVERSE MODEL. The inverse model iteratively adjusts the unknown parameter values to find a solution subject to the following constraints. Since the radar data contain errors, we do not want our inversion to fit these data exactly; we do not want to force the solution to simply minimize the mismatch between the data and the forward-model prediction. In addition, we expect the accumulation rate to vary smoothly along the flow band. Using a gradient method to solve this inverse problem [3, 4], these two independent constraints can be introduced to stabilize the solution process and to ensure that the solution is physically meaningful. This can be achieved by minimizing a performance index, \( I_p \):

\[
I_p = \|m\|^2 + v\|d\|^2 - T^2 \tag{1}
\]

The first term on the right is the model norm, the squared second spatial derivative of the accumulation rate integrated over the flow band. Penalizing large values of \(\|m\|^2\) prevents the solution from exhibiting roughness not required by the data. In the second term, \(\|d\|^2\) is the data norm, the sum of squared mismatches between the observations and the predictions.
normalized by the standard deviations of the data. The factor $\nu$ is a Lagrange multiplier, which is adjusted until the data norm equals a defined tolerance ($T$) that depends on the number of data. This value of $\nu$ sets the most appropriate trade-off between smoothness and fit.

**Results for Antarctica:** This method has been applied to a radar layer shown as the dotted black line in Figure 2, from Taylor Mouth, in Victoria Land, Antarctica [1]. The steady-state assumption is valid for layers at this depth; deeper layers could be used but would require a transient forward model.

In addition, we can explore the ability of this model to recover values of other unknown parameters. For Antarctica, we also solved for layer age. For Mars, we augment the parameter set with heat flux into the basal ice, and the ice-flow enhancement factor.

**Initial Results for the NPC:** Ice flow is assumed to be driven by surface-parallel shear stresses and the bed is assumed to be flat. Figure 4 shows internal layers generated by two prescribed accumulation-ablation regimes: (1) one zone of uniform accumulation and one zone of uniform ablation (result from [6]); (2) alternating zones of accumulation and ablation, known as accublation [7]. Other regimes will also be tested. These different internal structures retain some information about the accumulation-rate patterns that created them. Figure 4 shows internal layers for the two trial accumulation patterns. Regime (1) has an accumulation rate of 10 mm/yr and an ablation rate of 0.2 mm/yr. Regime (2) has six different accumulation and ablation zones, based on present-day trough locations. The accumulation rate is 1 mm/yr and the ablation rate is 0.2 mm/yr.

**Application to the Martian North Polar Cap:** In anticipation of future radar data along presumed Martian flow lines, we have generated synthetic internal layers that would result from different assumptions about ice-cap evolution and accumulation. Then, using our inverse method, we explore how well we can recover the accumulation patterns that created the layers. Assuming a two-stage history of NPC evolution, Winebrenner et al. [6] defined a relict pre-trough surface on Titania Lobe [5], the lobe adjacent to the main North Polar Cap (NPC). Adopting that surface geometry here, we construct flow lines by following the surface gradient. We use the forward model to generate internal-layers along these flow lines resulting from various accumulation patterns. We then attempt to recover the accumulation-rate pattern that generated the layer, when no additional information about surface velocity or layer age is available.

Synthetic internal layers, like those shown in Figure 4, can be used as “data” in our inverse procedure to explore the degree to which we can recover the original accumulation pattern, in preparation for analysis of actual radar data when they become available.

**References:**
INNER STRUCTURE OF THE POLAR LAYERED DEPOSITS ON MARS: RELATION TO THE CLIMATE RECORD. M. A. Kreslavsky\textsuperscript{1,2} and J. W. Head\textsuperscript{1}, \textsuperscript{1}Dept. Geological Sci., Brown University, Providence, RI, 02912-1846, USA, kreslavsky@brown.edu; \textsuperscript{2}Astronomical Institute, Kharkov National University, Kharkov, Ukraine.

Introduction: Polar layered deposits (PLD) on Mars have long been thought to contain a record of the past climate (see, e.g., \cite{1,2} and refs. therein). The inner structure and the evolution of the PLD is the subject of a number of controversial ideas and publications. Here we show that simple consideration of deposition and ablation of ice driven by any climate change produces many of the principal characteristics of the PLD. We focus on the north PLD (NPLD), and then consider applicability to the south PLD (SPLD).

Observations of layers: We systematically studied high-resolution MOC NA images of exposed layers in troughs at the highest latitudes. Apparent brightness alone is insufficient to identify layer sequences: it changes with illumination conditions and with seasons. We conclude that identification of the same layers should be done with high-resolution images on the basis of individual peculiarities of texture, topography, sequence and appearance; the same approach is proposed in \cite{3}.

Layer sequences exposed in the same troughs are remarkably identical over hundreds of kilometers \cite{2-4}. In the neighboring troughs, even in very close (tens of km) locations, the whole exposed sequences of layers are different, however, they contain remarkably identical segments within the sequences \cite{2,3}. Exposed unconformities inside the stack of layers are very rare \cite{1,2}, but there are a few examples (e.g., Fig. 1).

Interpretation: We consider a very simple scenario, in which the present-day NPLD is a result of a history of H$_2$O ice deposition and sublimation during some recent period of the geological history. Following \cite{1,2,5-7} we assume that a significant part of the NPLD was formed during the last ~5 Ma through global migration of H$_2$O from equatorial to high latitudes due to the general decrease of obliquity from 25-45\textdegree{} before 5 Ma ago to 15-35\textdegree{} during the recent 4 Ma. This particular time frame, however, is not essential for the following considerations. In some sense our scenario simultaneously simplifies and extends considerations from \cite{8} on the basis of modern data (topography, high-resolution images) and understanding of H$_2$O migration \cite{3-6}.

Simple deposition - ablation balance. Yearly amount frost accumulation and the yearly amount of sublimation are controlled by insolation regime (which is in turn controlled by the spin/orbital parameters) and by the atmospheric humidity, as well as by other secondary factors. There are reasons to assume (as a first approach) that the deposition - ablation balance (DAB) is a function of latitude and increases toward the pole (unless the obliquity is not too high). This means that at some latitude the balance is zero; net deposition (DAB>0) occurs poleward from this latitude and net ablation (DAB<0) occurs outside. This dividing latitude shifts back and forth due to climate change caused by (1) the change of the spin/orbit parameters ("astronomical forcing"), (2) availability of water vapor sources at lower latitudes (tropical mountain glaciers, high-latitude icy mantles, the opposite polar cap, groundwater discharge events), (3) internal climate instabilities. The outermost position of this boundary is well outside the present margins of the PLD; in the opposite extreme, the area of DAB=0 disappears, and the whole polar cap undergoes ablation. Through time such oscillations will produce a dome-shaped stack of deposits with a possible thin layer of deposits outside the dome and with a number of unconformities inside (Fig. 2). These unconformities will have an east-west oriented strike and a very shallow dip.

Albedo feedback. The DAB strongly depends on surface albedo. For brighter surfaces the balance is shifted toward deposition. On the other hand, it is natural to expect high albedo on the surfaces with net accumulation, where the surface is made of fresh small-crystal frost or snow, and lower albedo for the surfaces with net ablation, where old large-grain ice is exposed and residual impurities are accumulated. The present-day albedo dichotomy of the NPLD (very bright inner part and much darker troughs and some peripheral areas) probably maps the present-day positive and negative DAB. This mutual dependence of
DAB and albedo is a positive feedback. With this feedback, when the climate system goes through oscillations, the boundary latitude between positive and negative DAB will stay for some periods of time at its outermost and innermost position. (Fig. 3) This will result in steps in the generally domical shape of the NPLD. In the idealized model the steps would be concentric. When such steps are covered with later layers, unconformities with steeper dip, but still east-west strike are formed.

**Slope effect.** The DAB strongly depends on surface slopes. Tilted surfaces are exposed to a smaller portion of cold sky; in addition, at high latitudes, tilted surfaces, especially equator-facing ones, receive greater insolation (unless obliquity > ~40°). Thus, for tilted surfaces, the DAB is shifted toward ablation. This is consistent with the fact that we see lower albedo in the present-day troughs, especially on their equator-facing walls. This slope effect causes formation of troughs at the places where the steps were formed due to the albedo feedback. Fig. 4 shows topography of a part of the NPLD in grayscale. This image looks like shaded topography. This means that the present-day topography is a result of differentiation of some other, "original", step-like topography, in agreement with this scenario, except that in the real world the steps (and the troughs) are not perfectly concentric. The troughs eat away the steep-dip unconformities associated with the steps. The unconformity in the center of Fig. 1 fits the scenario very well: it is located near the trough axis, exposed at the trough end (and eaten away inside the trough); its character is a series of additional layers on the northern side of the trough.

**Conclusions.** We show that simple climate-controlled balance of deposition and ablation with albedo feedback and slope effect explains many characteristic properties of the NPLD, namely: (1) the general dome-shaped NPLD topography; (2) thin outliers; (3) troughs; (4) the general "shadow-like" character of their topography; (5) dome-like shape of the layers [2, 3]; (6) the identical section of layers within a single trough wall over long distances; (7) the absence of apparent unconformities on the trough walls over long distances; (8) non-identical sections of layers in neighboring troughs; (9) the presence of identical segments of the sections of layers in neighboring troughs; (10) the character and location of rarely observed unconformities; (11) the albedo pattern of the NPLD.

If our mechanism is correct, then the flow of the PLD material is minor, trough migration does not exceed a few km, and the climate record is readable in principle. Although each layer certainly corresponds to some time interval, it is unclear, what climate conditions control layer morphology. The total number of distinguishable layers well exceeds the number of alternating episodes of net deposition and ablation predicted by the spin/orbit oscillations [7].

Our simple considerations do not explain some other characteristic features of the NPLD: (1) the spiral trough pattern; (2) the typical slope of trough walls and its latitudinal trend, (3) Chasma Boreale and similar features. Some additional processes and/or effects must be considered to account for them.

All our considerations are generally applicable to the SPLD. Obvious differences (and complications) for SPLD are (1) the presence of a local thin CO₂ ice unit, (2) the lower albedo indicating a negative DAB for most of the SPLD, (3) an older crater retention age indicating generally lower deposition over some recent period, (4) a probably larger role of flow, (5) other denudation mechanisms (e.g., eolian erosion) in addition to sublimation.