Geophysical Research Letters

RESEARCH LETTER
10.1002/2016GL068457

Stratigraphy and evolution of the buried CO$_2$ deposit in the Martian south polar cap

C. J. Bierson$^1$, R. J. Phillips$^{2,3}$, I. B. Smith$^4$, S. E. Wood$^5$, N. E. Putzig$^6$, D. Nunes$^6$, and S. Byrne$^7$

$^1$Department of Earth and Planetary Sciences, University of California, Santa Cruz, California, USA, $^2$Planetary Science Directorate, Southwest Research Institute, Boulder, Colorado, USA, $^3$McDonnell Center for the Space Sciences and Department of Earth and Planetary Sciences, Washington University, St. Louis, Missouri, USA, $^3$Department of Space Studies, Southwestern Research Institute, Boulder, Colorado, USA, $^5$Department of Earth and Space Sciences, University of Washington, Seattle, Washington, USA, $^6$California Institute of Technology, Jet Propulsion Laboratory, Pasadena, California, USA, $^7$Lunar and Planetary Lab, University of Arizona, Tucson, Arizona, USA

Abstract Observations by the Shallow Radar instrument on Mars Reconnaissance Orbiter reveal several deposits of buried CO$_2$ ice within the south polar layered deposits. Here we present mapping that demonstrates this unit is 18% larger than previously estimated, containing enough mass to double the atmospheric pressure on Mars if sublimated. We find three distinct subunits of CO$_2$ ice, each capped by a thin (10–60 m) bounding layer (BL). Multiple lines of evidence suggest that each BL is dominated by water ice. We model the history of CO$_2$ accumulation at the poles based on obliquity and insolation variability during the last 1 Myr assuming a total mass budget consisting of the current atmosphere and the sequestered ice. Our model predicts that CO$_2$ ice has accumulated over large areas several times during that period, in agreement with the radar findings of multiple periods of accumulation.

1. Introduction

Obliquity and orbital parameters strongly force the insolation and thus stability of both CO$_2$ and water ices at different latitudes on Mars [Murray et al., 1973; Ward, 1973, 1974; Laskar et al., 2002, 2004; Manning et al., 2006; Levrard et al., 2007]. In the current epoch, surface water ice is most stable at the poles, and there is strong evidence that water cycles between midlatitudes at high obliquity and the poles at low obliquity [Kreslavsky and Head, 2002; Head et al., 2003; Madeleine et al., 2009].

Carbon dioxide ice is less stable at the surface of Mars than water ice and undergoes strong seasonal cycles of deposition and sublimation. These seasonal cycles drive the transport of large quantities of CO$_2$ between the north and south pole each year, which in turn has a strong seasonal effect on the surface pressure of the thin atmosphere (∼600 ± 100 Pa) [Hess et al., 1979, 1980; Tillman et al., 1993].

Longer cycles also exist for CO$_2$ deposits at the South Pole. The South Pole Residual Cap (SPRC) is a several meters thick veneer composed primarily of CO$_2$ ice. Recent work has shown that the SPRC is sublimating in some places and accumulating in others [Thomas et al., 2013, 2016; Becerra et al., 2015], with a possible cycle time for the residual CO$_2$ cap of about 100 years [Bryne et al., 2014]. Corresponding to even longer cycles, CO$_2$ ice is found in reservoirs other than the atmosphere and thin surface deposits. Using Shallow Radar (SHARAD) instrument data [Seu et al., 2007] from the Mars Reconnaissance Orbiter (MRO), Phillips et al. [2011] mapped a massive deposit of buried CO$_2$ ice within the south polar layered deposits (SPLD) in a radar stratigraphic unit called “reflection-free zone three” (RFZ$_3$). They measured volumes that if sublimated would increase the surface pressure of Mars by 65–85% (400 to 500 Pa). Their results support climate models that predicted partial atmospheric collapse on the poles over cycles of the planet’s obliquity [Ward et al., 1974; Toon et al., 1980; François et al., 1990; Jakosky et al., 1995; Forget, 1998; Armstrong et al., 2004; Manning et al., 2006; Wood et al., 2012].

1.1. Overview

Many groups have characterized aspects of Mars’ south polar cap, and various naming schemes for different units have been developed [Tanaka et al., 2007; Thomas et al., 2013; Phillips et al., 2011]. This paper will use a naming scheme modified from that of Tanaka et al. [2007] and given in Table 1. The compositional
Table 1. Stratigraphic Units in the Australe Mensa Region of the SPLDb

<table>
<thead>
<tr>
<th>Group Names</th>
<th>Unit Name</th>
<th>Ice Composition</th>
<th>Unit Thickness (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>SPRC</td>
<td>A4b</td>
<td>CO₂</td>
<td>10b</td>
</tr>
<tr>
<td></td>
<td>A4a</td>
<td>H₂O</td>
<td></td>
</tr>
<tr>
<td>RFZ₃</td>
<td>BL₃</td>
<td>H₂O</td>
<td>&lt; 20</td>
</tr>
<tr>
<td></td>
<td>AA₃c</td>
<td>CO₂</td>
<td>300</td>
</tr>
<tr>
<td></td>
<td>BL₂</td>
<td>H₂O</td>
<td>40</td>
</tr>
<tr>
<td></td>
<td>AA₃b</td>
<td>CO₂</td>
<td>&lt; 700</td>
</tr>
<tr>
<td></td>
<td>BL₁</td>
<td>H₂O</td>
<td>20</td>
</tr>
<tr>
<td></td>
<td>AA₃a</td>
<td>CO₂</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td>AA₂</td>
<td>H₂O</td>
<td>300c</td>
</tr>
<tr>
<td></td>
<td>AA₁</td>
<td>H₂O</td>
<td>3500c</td>
</tr>
</tbody>
</table>

a Units are ordered with the surface at the top. Methods used to determine ice compositions and thickness in RFZ₃ are presented in sections 2 and 3. Note that the geologic unit AA₃ is not subdivided in Tanaka et al. [2007].

b Thomas et al. [2005] and Tanaka et al. [2007].
c Kolb et al. [2006].
labels assigned to each unit are justified in sections 2 and 3. The radar unit RFZ₃ [Phillips et al., 2011] and observationally mapped unit AA₃ will be used interchangeably as justified in section 2.

With an enhanced radar mapping effort, we show that the RFZ₃ unit is more extensive and locally deeper than originally mapped. We find three distinct CO₂ subunits, each capped by a bounding layer (BL). We use a combination of radar and surface observations to constrain the composition of the observed subunits. We propose a scenario in which CO₂ ice is deposited over much of the poles during low obliquity periods. This ice subsequently retreats until a remnant is sequestered below a water ice deposit (BL subunits), removing it from contact with the atmosphere.

2. Radar Observations

Phillips et al. [2011] mapped four reflection-free zones (RFZs) within the SPLD. Based on estimates of permittivity, the most poleward of those zones, RFZ₃, was shown to be massive CO₂ ice with a volume of ~4500 km³. This volume incorporated locations where RFZ₃ was easily detected with SHARAD but excluded the region poleward of 87°S, where there is no radar coverage due to the orbital inclination of MRO. Their analysis included observations from 129 orbital passes, and the coverage was too sparse to find the northern boundary of the deposits or completely fill gaps. They demonstrated a reasonable but imperfect correlation of RFZ₃ with the image-based geologic unit AA₃ [Kolb et al., 2006; Tanaka et al., 2007; Phillips et al., 2011]. To provide a better estimate of the total volume of RFZ₃, the measured unit was extrapolated to coincide with the boundaries of AA₃ as mapped in optical imagery.

Here we improve upon the prior study by including an additional 300 radar observations that cover a larger area and fill gaps in the previous coverage. With this new mapping effort, detections of RFZ₃ correspond to the northern bounds of the optically imaged AA₃ (Figure 1). We also find the unit in two previously unmapped areas and include a portion of AA₃ omitted from the previous study (red arrows in Figure 1). This improvement in correlation between AA₃ and RFZ₃ carries a twofold importance. First, it provides confidence in using AA₃ to extrapolate the extent of RFZ₃ poleward of 87°S. Second, it strengthens the case for interpretations of RFZ₃ as a CO₂ deposit based on optically imaged sublimation troughs and pits that are unique to AA₃. No other unit on the water ice dominated SPLD exhibits these features [Phillips et al., 2011].

Further improvements are provided by better detection of the RFZ₃ base (Figure 2). At the most southerly latitudes observed by SHARAD, our mapping yields maximum depths of just over 1000 m (blue arrow in Figure 1), whereas ~700 m was the previously reported maximum. Better spatial coverage and enhanced radar processing have improved confidence in the placement of this lower boundary.
Figure 1. (a) Maximum depth map of RFZ₃ deposits and (b) extent of the bounding layers. In Figure 1b BL₁ is mapped as orange and BL₂ is mapped in blue. Base geologic map is from Tanaka et al. [2007]. Red arrows indicate areas not mapped by Phillips et al. [2011]. The location of the thickest (>1000 m) part of RFZ₃ mapped is indicated by the blue arrow. The green arrow points to a location that was found to be thicker than was estimated in Phillips et al. [2011] due to the presence of a deeper bounding layer 1 (BL₁).

Another enhancement of our depth measurements is found at lower elevations. In addition to the original bounding layer (BL₂) that separates two CO₂ subunits of RFZ₃ in many locations, we find a second bounding layer (BL₁) (Figure 2). Where it is present, BL₁ was previously interpreted to be the base of RFZ₃. The inclusion of the third subunit increases the measured depth of the RFZ₃ base in this location by as much as 300 m (green arrow in Figure 1).

BL₁ and BL₂ are not exposed on the surface, so constraints on their material properties are limited to radar measurements. The strong reflection produced at the boundaries requires that the BLs must have permittivity distinct from the surrounding CO₂. Volumetrically, the only viable candidate material is water ice. This water ice may, however, have some dust and CO₂ (as a clathrate hydrate) incorporated into it [Jakosky et al., 1995]. Using a forward model (section S1), we find the concentration of such contaminants must be low. When considering a permittivity characteristic of water ice (3.15) the bounding layers vary in thickness from less than the resolution limit of SHARAD (~11 m in water ice) to 64 m with a mean of 37 m.
The expanded SHARAD coverage and the newly measured depths add to the measured volume of RFZ3, which has increased from 4500 km$^3$ to 7700 km$^3$ prior to extrapolation (Figure 1a). We extrapolate poleward bounded by the optically imaged extent of AA3 using a natural neighbor method and obtain a new volume of 14,800 km$^3$ for RFZ3, 18% greater than that of Phillips et al. [2011]. We assume zero porosity and no contaminants, so the density of RFZ3 is that of CO$_2$ ice, $\sim$1600 kg/m$^3$ [Piqueux et al., 2003], and the total mass of this unit is then $2.4 \times 10^{16}$ kg, essentially equal to the mass of the Martian atmosphere ($2.5 \times 10^{16}$ kg) [James and North, 1982]. If totally sublimated, this mass would increase the current average surface pressure by 610 Pa, doubling the present value and enhancing the stability of liquid water at the surface.

3. Surface Observations

Figure 3 presents a stretched CTX image of a part of the SPLD containing the SPRC CO$_2$ and H$_2$O (AA$_4a$ and AA$_4b$) subunits. Both are bright, but the CO$_2$ has a mottled appearance due to the "Swiss cheese" sublimation features [Thomas et al., 2005]. The unit AA$_3$ is darker and uniquely contains troughs (not shown) that exhibit a mixture of dust, water ice, and dry ice, as measured by the OMEGA instrument in late summer [Douté et al., 2007]. Thus, spectral measurements do not provide compelling evidence for or against a water ice cap over RFZ$_3$ (BL$_3$). However, a polygonal pattern (Figure 3b) at the surface of AA$_3$ with a characteristic dimension of about 20 m is found on no other unit in the SPLD. These polygons are highly distinct from the CO$_2$ sublimation features on the SPRC and most closely resemble polygonal features seen in other thermally stressed water ice deposits on Mars [Mellon et al., 2008]. Thus, we propose that the polygons are unique to AA$_3$ because it is the only region in the SPLD where dry ice lies beneath a relatively thin layer of water ice. Given this, the observed polygonal terrain could be caused by stresses initiated by the fact that the coefficient of thermal expansion of CO$_2$ ice [Manzhelii et al., 1971] is an order of magnitude larger than that for water ice [Röttger et al., 1994]. If BL$_3$ is 20 m thick SHARAD would resolve a basal reflection (which it does not), and the thermal stress would decrease by about a factor of 2.5. Therefore, BL$_3$ must be thinner than BL$_2$ on average.

If both BL$_3$ and AA$_{4b}$ are water ice layers, it raises a question of why they are so distinct in observations. Figure 3 shows that BL$_3$ is noticeably darker than AA$_{4b}$ and only BL$_3$ shows the polygonal terrain. One hypothesis is that these layers were deposited with different dust fractions. Alternatively, BL$_3$ may have formed with the same initial dust fraction as AA$_{4b}$ and subsequent sublimation of BL$_3$ produced a dark lag deposit at the surface. Both scenarios require two periods of water ice deposition. Modeling by Montmessin et al. [2007] suggests the last period of water ice deposition at the south polar cap was ~20 kyr ago with their largest uncertainty being how dust cycles change with orbital parameters. Future modeling efforts should investigate if variable dust loading or other factors could cause the difference observed in these two units.

4. Modeling CO$_2$ Ice Accumulation

To model the depositional history of CO$_2$ ice at the poles, we followed the methods of Manning et al. [2006] and Wood et al. [2012], who performed seasonally resolved calculations of the evolution of Mars’ atmospheric pressure and polar CO$_2$ deposits from 800 kyr to the present. This zonally symmetric model includes subsurface
heat conduction, insolation-dependent albedo [Guo et al., 2010], and Mars orbital variations [Laskar et al., 2004]. In its base setup this model has surface CO₂ in constant contact with the atmosphere. For more details on the model, see Wood et al. [2012] and supporting information.

Our model predicts that the SPLD has experienced five periods of CO₂ accumulation up to 150 m at 89°S during the last 800 kyr (Figures 4 and S4). Each period of CO₂ ice deposition corresponds to the removal of a few hundred pascals from the atmosphere (Figure 4). Our base model does not sequester the ice, and it returns to the atmosphere at the end of each period of high obliquity (Figure S4). However, the presence of the RFZ₃ unit requires some mechanism to stabilize and protect the deposit in periods of high obliquity. Given the evidence of water ice bounding layers (see previous section), it is natural to assume that water ice stabilizes each RFZ₃ subunit during periods of high obliquity.

The amount of water ice needed to thermally protect an underlying CO₂ deposit can be estimated by comparing the CO₂ phase curve with an estimate of the annual thermal wave. Assuming a single-mode thermal wave [Turcotte and Schubert, 2014], we calculate the maximum temperature at a given depth ($T_{\text{max}}$) as

$$T_{\text{max}} = T_{\text{mean}} + \Delta T \exp(-z/d_{\text{w}}),$$

where $T_{\text{mean}}$ is the mean annual surface temperature, $\Delta T$ is the annual temperature anomaly, $z$ is the depth below the surface, and $d_{\text{w}}$ is the annual thermal skin depth. This single-mode assumption does not capture the fact that during the winter the surface temperatures are being fixed at the sublimation temperature of CO₂. To try and offset this effect, we over estimate $\Delta T$. Our modeled results indicate that approximately 10 m of water ice is enough to stabilize the CO₂ ice even with a large thermal wave (Figure S2). This result is consistent with previous work by Jakosky et al. [1995]. The thickness of a water ice layer needed to stabilize the underlying CO₂ is not very sensitive to $\Delta T$ because of the exponential decay of the thermal wave with depth. It is, however, very sensitive to $T_{\text{mean}}$, which is itself a function of obliquity. Taken at face value, this model implies that 37 m of H₂O ice could stabilize the underlying CO₂ ice up to a mean annual temperature of ~190 K. Even at large obliquity, $T_{\text{mean}}$ is not be expected to rise above ~180 K [Jakosky et al., 1995]. This implies that the stability of the AA₃ subunits is controlled by the stability of the water ice BL’s capping them.

This optimistic preservation model assumes that the CO₂ is under lithostatic pressure and thus requires that no cracks penetrate through the overlying water ice. This assumption is validated by the fact that CO₂ sublimation pits in AA₃ are rare unless they are associated with troughs, where they are plentiful [Phillips et al., 2011]. We attribute this association to the local relief of lithostatic pressure in the troughs.

To estimate a minimum age for AA₃, we incorporate the insulating effect of the BLs during periods of high obliquity. This is done by removing the CO₂ ice from contact with the atmosphere in some climate model runs, effectively sequestering it over several obliquity cycles. This step allows the model to predict the evolution of surface pressure and total thickness of the CO₂ ice deposits integrated over 380 kyr (Figures 4 and S5). These simulations reveal that during certain periods, the surface pressure regularly drops below the current value of 610 Pa and that its long-term equilibrium value is ~1250 Pa. Annual variability in surface pressure may range from 1100 Pa to 1350 Pa during these periods (Figure S4).
5. Discussion

Our improved mapping of RFZ and modeling work presented above provide a framework to understand the history of CO₂ reserves on Mars. Previous studies had shown that large quantities of CO₂ can be deposited on the polar cap during times of low obliquity [Armstrong et al., 2004; Manning et al., 2006; Wood et al., 2012]; however, those studies (with the exception of Wood et al. [2012]) underestimated the CO₂ budget of the planet because they did not incorporate the full quantity of CO₂ presently sequestered in the SPLD. Furthermore, those studies did not include mechanisms by which the deposits might persist between periods of low obliquity.

The radar data shows that the ice distribution is not symmetric about the pole but is longitudinally bound between ∼240° and ∼360°E, a pattern that is also consistent with the residual CO₂ cap. There are three,
nonexclusive, mechanisms that might explain this distribution, a bias in preservation: a bias in accumulation, or lateral movement. The area where the CO$_2$ resides is a local minimum in the topography. This could cause a preservation bias if it allows the overlying water ice to more easily remove the CO$_2$ from contact with the atmosphere. Lower altitudes are also favored for CO$_2$ deposition because of the higher atmospheric pressure [Kreslavsky and Head, 2005]. Global circulation models have demonstrated the Hellas and Argyre basins have pronounced effects on where CO$_2$ ice is deposited [Colaprete et al., 2005] and similar forcing may have been present for the deposition of AA$_3$. Lastly, there is some literature discussing the geologic evidence for CO$_2$ glaciers on Mars [Kreslavsky and Head, 2011]. While evidence for CO$_2$ glaciers has not been observed in the Southern hemisphere [Kreslavsky and Head, 2011], we cannot rule out the possibility that the areal distribution of CO$_2$ has been affected by glacial flow. Future work should examine each of these possibilities in more detail to better understand the history of Martian CO$_2$ transport.

We find evidence of two bounding layers between the CO$_2$ subunits of RFZ$_3$ suggesting that the CO$_2$ ice was emplaced during three periods. A bounding layer caps each RFZ$_3$ subunit and protects it during periods of high obliquity. Our measurements of the dielectric properties of the bounding layers are consistent with nearly pure water ice between 15 and 60 m thick. Modeling described in section 4 finds that these thicknesses are sufficient for reducing the sublimation rate of CO$_2$ ice to near-zero values.

This opens the question of what controls the change from CO$_2$ deposition to water ice deposition. Montmessin et al. [2007] have suggested that in the recent past, changes in the longitude of perihelion may control which polar cap is favorable for water ice deposition. The thickness of the BLs could be accumulated in tens of thousands of years at the deposition rates suggested by Montmessin et al. [2007]. The longitude of perihelion changes much faster than obliquity, with a period of about 50 kyr [Laskar et al., 2004]. While the details should be studied carefully by future work, it is at least plausible that these changing orbital parameters could switch between regimes of CO$_2$ deposition to water ice deposition.

Our model predicts that during the last 1000 kyr, four periods of perennial CO$_2$ ice deposition left thick, extensive CO$_2$ deposits at observed RFZ$_3$ latitudes (84°S–87°S). We suggest that during subsequent sublimation, three of those deposits were preserved in topographic depressions by water ice layers, effectively halting the loss of CO$_2$ reserves into the atmosphere. Each time, the atmospheric pressure returned to lower values than the long-term equilibrium. During the periods of greatest collapse (which occurred on both poles), the surface pressure may fall to values as low as 200 Pa, about a third of the present-day value. These periods profoundly impede the mobility of dust, sand, and volatiles. The water ice polar caps, covered by CO$_2$ ice, may be cut off entirely from the atmosphere, reducing volatile availability via another mechanism. Finally, at the highest pressures expected during this period, 1350 Pa, liquid water would be stable at many more locations that at present, and dunes that appear inactive in the current climate could become active, explaining the divergent orientations of many dunes relative to current wind regimes [Ewing et al., 2010]. This proposed history provides the minimum possible age of the AA$_3$ deposit, ~300 kyr for the oldest units. It is possible that older units may have been preserved in place of younger ones, a question that should be explored by future models of CO$_2$ and water ice stability on obliquity timescales.

Acknowledgments
Data used in this work are available from the NASA Planetary Data System or from the Colorado SHARAD Processing System (CO-SHARPS). E-mail co-sharps@spaceops.swri.org for more information. The authors thank the two anonymous reviewers for their feedback.

References
Geophysical Research Letters


Tanaka, K., E. Kolb, and C. Fortezzo (2007), Recent advances in the stratigraphy of the polar regions of Mars, LPI Contributions, 3–6.


