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Key Points:

- Radar reflections can be used to constrain the dust content of individual layers in the north polar layered deposits of Mars
- Layer dust contents are higher than predicted by models, indicating a gap in our understanding of how the global dust cycle varies with time
- Dust content and completeness of the climate record vary strongly by location and with depth

Supporting Information:

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Radar Reflectivity as a Proxy for the Dust Content of Individual Layers in the Martian North Polar Layered Deposits

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Abstract The stratigraphy of the north polar layered deposits (NPLD) of Mars is believed to contain a climate record of the recent Amazonian period. However, full utilization of this record is difficult without detailed information regarding the physical properties of the constituent layers. Here we present a method for determining the fractional dust content of individual layers using a combination of orbital radar reflectivity measurements and physical modeling. We apply this method to the upper 500 m of the NPLD at 10 study sites and compare the results to a cap-wide radar-mapped surface. Our results show that reflectivity can vary drastically both geographically and with depth, a result we attribute to changing dust content, though the impact of variable layer thickness cannot be totally discounted. These findings imply large-scale regional patterns in ice and dust accumulation do not remain consistent through time. We also find that current models of Mars's dust cycle and polar ice accumulation consistently underpredict the dust content of layers, indicating that our understanding of dust transport, dust sequestration, or dust preservation remains incomplete. Comparisons of study sites on the NPLD also show that some locations contain fewer radar reflectors than others, meaning they may contain a less complete record of the planet's recent paleoclimate, and any future efforts to use the polar layered deposits as a climate proxy, including in situ measurements, should take this into account by choosing study sites wisely.

Plain Language Summary The north polar cap of Mars is very similar to ice sheets on Earth. On Earth, scientists use the layers of ice to learn about how the climate has changed over time. It should be possible to do something similar on Mars, but we cannot drill ice cores on Mars, and so far, scientists have had difficulty pulling useful information out of the layers. In this paper, we show that it is possible to use radar reflections in the ice to get a rough idea of how much dust is in a certain type of layer. The brightness of a reflection is related to the composition of the layer that caused it, so after making a few assumptions, we can first measure that brightness and then use a model to estimate how much dust would have to be present in the ice to get a matching reflection. After doing this for many reflections at 10 different study sites, we found that some layers are a lot dustier than we previously believed, meaning our previous understanding of the Mars ice and dust cycles is incomplete. We also found that the composition of layers was very different from place to place, which means local conditions are playing a large role in ice and dust accumulation.

1. Introduction

The north polar layered deposits (NPLD) of Mars contain the second largest surface reservoir of water ice on the planet, covering a quasi-circular region ~1,000 km in diameter and up to 2.0 km thick (Phillips et al., 2008; Tanaka, 2005). Thousands of subparallel layers believed to be composed of ice and dust are visible in optical imagery, giving the formation its name, and radar sounding data reveal tens of reflectors that parallel the visible layers (Christian et al., 2013; Phillips et al., 2008). Analogous to layering in polar ice on Earth, the NPLD are believed to record information on Mars's climate (e.g., Cutts & Lewis, 1982; Laskar et al., 2002; Toon et al., 1980). Based on modeling of the stability of ice at polar latitudes, the oldest layers of the NPLD are usually thought to be roughly four million years old (Greve et al., 2010; Levrard et al., 2007), though some maintain that the lower NPLD may be much older (Tanaka et al., 2008). Evidence suggests the uppermost surface is extremely young and possibly in net positive accumulation in the present day (Brown et al., 2016; Herkenhoff & Plaut, 2000; Landis et al., 2016; Tanaka, 2005). Because NPLD accumulation of ice

and dust is affected by many global processes, the polar cap should record valuable information about how the Martian climate has evolved in the recent past.

Efforts aiming to link layering or reflector geometry to climate processes have produced mixed results. Stratigraphic sections visible in outcrops have been shown to contain cyclic bedding features (Becerra et al., 2016, 2017; Hvidberg et al., 2012; Milkovich et al., 2008; Milkovich & Head, 2005; Perron & Huybers, 2009), but the period and scale of these features is nonuniform across the cap, and connecting them to climate conditions remains a challenge (Becerra et al., 2016; Hvidberg et al., 2012). Some success has been found in correlating Mars's modeled orbital dynamics (Laskar et al., 2002; Levrard et al., 2007) to large packets of radar reflectors (Hvidberg et al., 2012; Phillips et al., 2008; Putzig et al., 2009) and an unconformity in the upper section of the NPLD (Smith et al., 2016). However, these correlations are often inconsistent and are too broad to enable the use of radar reflectors as a detailed climate record. Recent work has suggested that these radar reflectors may share a genetic link with the outcrop stratigraphy, and indeed, some research has pointed to specific layers known as "marker beds" as a plausible source for radar reflectors (Christian et al., 2013; Lalic & Holt, 2016), though so far, such a connection is largely an assumption.

In an effort to fill these knowledge gaps and clarify the relationship between NPLD layering and the Martian climate, we have measured reflection power and derived reflectivity values for a sequence of reflectors at 10 sites across the NPLD. Combining these reflectivities with a thin layer reflection model based on the one presented in Lalic and Holt (2016) allows us to infer the relative dust content of individual layers. Later, we compare the calculated dust content to the concentration predicted by a previously published model of NPLD accumulation. That model used Mars's orbital parameters to predict how the accumulation rates of ice and dust varied in the relatively recent past at the planet's north pole (Hvidberg et al., 2012).

2. Background

2.1. Outcrop Stratigraphy

Malin and Edgett (2001) were the first to define the original "marker bed" and identify it in multiple images across different outcrops (Figure 1). Soon after, others were able to more definitively correlate sequences containing this original marker bed, and others like it throughout the NPLD (Becerra et al., 2016; Fishbaugh et al., 2010; Fishbaugh & Hvidberg, 2006). Laskar et al. (2002) and Milkovich and Head (2005) found that these sequences contained a periodic signal, with a periodicity of ~20–30 m. This behavior was thought to represent changes in dust and ice accumulation rates driven by Mars's insolation cycle. Perron and Huybers (2009) failed to detect this same behavior at a statistically significant level but were able to identify a signal at ~1.6-m spatial wavelength.

Later work showed that albedo profiles are prone to obfuscation by surface processes or imaging conditions (Fishbaugh et al., 2010; Herkenhoff et al., 2007), reducing the certainty in interpretation. To help, high-resolution topography was suggested as a supplemental means of dividing layers based on intrinsic physical properties (Fishbaugh et al., 2010). Becerra et al. (2016) showed that sequences of ablation-resistant layers (determined to be the same marker beds as traced before) identified using high-resolution digital terrain models in combination with imagery are indeed more easily correlated with each other across the NPLD than brightness profiles alone. Further investigation showed that these sequences contain dominant periodicities consistent with those found in the modeled insolation record (Becerra et al., 2017), which supports the interpretation that marker beds are dust rich layers as suggested by Hvidberg et al. (2012). This connection remains tenuous, however, as errors in periodicity measurements are large and identification of signals is inconsistent from outcrop to outcrop. Furthermore, while identifying such a connection between visible layering periodicity and insolation cycles represents an important step forward, it is difficult to use these techniques to constrain individual layer properties.

2.2. Radar Stratigraphy

In addition to the work done with outcrop stratigraphy, several attempts have been made to tie radar reflectors detected in data collected by the Shallow Radar (SHARAD) orbital radar sounder to the Martian paleoclimate. In general, radar reflections are caused by abrupt changes in average subsurface permittivity. In the NPLD, it is assumed that reflections are the result of discrete layers of ice enriched in dust relative to the material around them or as packets of enriched layers below the resolution of the radar (Nunes &

Figure 1. Portion of High Resolution Imaging Science Experiment image PSP_009967_2660 (85.86°N, 102.12°E) showing a section of the upper 500 m of the north polar layered deposits. The arrow indicates the original marker bed identified by Malin and Edgett (2001). The outcrop slopes toward the southwest or the bottom left of the image. Illumination is from the southwest.

Phillips, 2006). This enrichment must be caused by a variable ratio of dust to ice during accumulation or by preferential loss of material (e.g., sublimation) subsequent to deposition, reflecting the local and global climatic states at the time. Several forcing scenarios have been proposed based on obliquity or perihelion (Hvidberg et al., 2012; Phillips et al., 2008; Putzig et al., 2009), but definitive correlations between forcings and observed layering have not been made (Figure 1).

Phillips et al. (2008) identified four discrete packets of reflectors in the NPLD separated by radar-dark “interpacket” zones of presumably more homogeneous material (Figure 2). They interpreted this large-scale structure as the result of periodic changes in ice and dust accumulation due to variations in Mars’s eccentricity and obliquity; however, a later study found that several outcrops of interpacket zones have no visibly obvious differences from packet zones (Smith & Holt, 2015). Further work by Putzig et al. (2009) modified this interpretation by correlating reflector packets to low-amplitude variations in the insolation function, rather than obliquity (the primary driver of polar insolation), and presented the hypothesis that reflector-causing layers are formed by fluctuations in dust content during periods of relatively constant ice deposition, similar

to the mechanism originally proposed by Laskar et al. (2002). This interpretation is at odds with previous work by Levrard et al. (2007), who hypothesized the formation of dust-rich layers through sublimation lag deposits during periods of high obliquity. This ambiguity has yet to be resolved.

More recent work attempted to establish correlations between imagery-based outcrop stratigraphy and radar stratigraphy. Christian et al. (2013) applied signal analysis techniques to radar traces near a trough wall and found that while some peak wavelengths were shared between the radar data and the nearby stratigraphic profile, many were not. Still, this suggests some commonality between the two types of observations (separated by an 8 times factor of resolution). Lalach and Holt (2016) compared reflectivities of three spatially mapped SHARAD reflectors to those predicted by a model approximating reflection from dust rich layers similar to marker beds, and while they found that the model was capable of accurately reproducing the observations, the solutions were nonunique, and they were unable to place tight constraints on layer properties. They also found that radar reflectivity varied greatly both within and between the mapped reflectors, as expected (Phillips et al., 2008). They attributed both forms of variation primarily to differences in layer thickness, proposing that radar reflectivity could potentially be used as a proxy for relative deposition rates.

Figure 2. Shallow Radar radargram 1716901000 crossing the north polar layered deposits (NPLD). Inset is a Mars Orbiter Laser Altimeter colorized elevation map of the NPLD containing the radargram ground track. The vertical axis is in time delay, rather than depth. Typical NPLD thickness is approximately 1.5 km. The reflector packets delineated by Phillips et al. (2008) are labeled 1–4. Only packet 1 was analyzed here.

2.3. Layer Formation Processes and Climate Modeling

In general, it is assumed that the composition of NPLD layers is determined by changes in the relative rates of net dust and ice accumulation (Hvidberg et al., 2012). As previously discussed, two processes have been proposed that could potentially form layers with high enough dielectric contrasts to cause radar reflections (Lalich & Holt, 2016). The first process involves the accumulation and subsequent sublimation of ice from a dusty ice layer. Dust impurities present in the layer will be left behind when the ice sublimates, forming a more pure dust layer on the surface, with relatively low interstitial ice content. The ice acts only as a cementing agent. Once enough dust builds up on the surface, it blocks vapor diffusion from below and preserves the ice underneath (Bramson et al., 2017; Paige, 1992; Skorov et al., 2001). Exactly how much dust is required to create such a deposit is a matter of debate but estimates place the layer thickness necessary to fully protect the underlying ice between a few millimeters (Skorov et al., 2001) and a few centimeters (Paige, 1992) in the polar regions. Modeling has shown that when the dust layer reaches a few centimeters, sublimation is reduced to effectively zero (Bramson et al., 2017; Bramson et al., 2019), indicating that dust lags should never reach 10 cm thick.

The second suggested process for creating polar layers involves smaller fluctuations in the relative deposition rates of ice and dust. Rather than forming during periods of ice removal, in this scenario reflector-causing layers are the result of constructional processes, formed when climate conditions facilitate significant deviation in average ice or dust deposition rates (Cutts & Lewis, 1982; Laskar et al., 2002). Stochastic processes such as volcanic ash or crater ejecta falling on the NPLD may also have created individual or spatially local layers, but it is unlikely that they are responsible for the bulk of the widespread, regularly spaced reflectors. The lag and accumulation hypotheses are not mutually exclusive, but they are expected to be active during opposite periods in the planet's orbital cycle (Hvidberg et al., 2012), and if both are creating reflector-causing NPLD layers, then it becomes difficult to explain the large low reflection packets present within the stack of layers (Putzig et al., 2009).

Hvidberg et al. (2012) modeled NPLD accumulation using ice and dust deposition rates forced by a temperature gradient between the poles and the equator. This, in turn, was driven by the planetary obliquity cycle. Their model allowed for creation of layers using both processes discussed above. Rather than dividing layers into lag deposits and constructional layers, they instead observed a dichotomy between layers formed during periods of high dust deposition and those formed during periods of low ice deposition, of which lag deposits are a subset. Broadly speaking, the model predicted these layering processes to occur 180° out of phase in the obliquity cycle, resulting in an alternating pattern of “high dust deposition” and “low ice deposition” layers separating periods of “average” ice accumulation. These two types of layer-forming mechanisms produced distinct ranges of dust contents. Layers formed during peak dust deposition periods typically had less than 5% dust content, while those formed during periods of minimum ice deposition often had dust contents above 5%. Modeled high dust content layers were also found to be concentrated in packets, similar to the radar reflectors observed by SHARAD. This finding suggested that radar reflectors are the result of layers formed when ice accumulation is lowest and in some cases possibly by lag deposits.

3. Data and Methods

3.1. SHARAD Data and Study Sites

The SHARAD instrument is a radar sounder on the Mars Reconnaissance Orbiter. It uses a chirped signal with a bandwidth of 10 MHz at a 20-MHz center frequency transmitted in an 85- μ s pulse (Seu et al., 2007). SHARAD's nominal range resolution is 8.4 m in water ice, though in practice this can approach 20 m in the NPLD (Nunes & Phillips, 2006; Seu et al., 2007). Cross-track resolution is between 3 and 6 km, while an along-track resolution of 0.3–1 km is achieved using synthetic aperture processing (Seu et al., 2007).

In order to select sites for the extraction of radar profiles, we used a previously mapped cap-wide radar horizon hypothesized to correspond to a global climatic change (Smith et al., 2016; Figures 3 and 4). Depending on location, this horizon may manifest as an unconformity in places where material was removed, or a lag or disconformity, when large-scale ablation is not observed. Ten sites with widely varying local reflectivities were identified and selected using the cap-wide reflector. Many of the sites were chosen in regions of the NPLD that are difficult to link directly to the other sites using traditional means, due to the disconnection of layering across troughs. Methods to connect these discontinuous sections are described in Smith and Holt (2015).

Figure 3. (a, b) Uninterpreted Shallow Radar radargrams of the north polar layered deposits. (c, d) The same radargrams showing the cap-wide radar horizon used to select and compare study sites in blue. Arrows indicate which direction is north. (e) Mosaic of the north polar region created using Viking imagery collected between 1976 and 1980 (McEwen et al., 1994) converted to gray scale, with ground tracks of radargrams shown in color.

Three sites (A, B, and E) were chosen in the “saddle region” of the NPLD, which connects the main lobe of the polar cap to the Gemina Lingula formation. These three sites are linked via continuous subsurface reflectors (Lalich & Holt, 2016), confirming that any profiles extracted from these sites will sample the same subsurface layer sequence. All sites were also chosen with a preference for dense radar coverage.

3.2. Measuring Radar Reflectivity

In order to facilitate comparisons of radar reflectivity with modeled paleoclimate conditions, profiles of radar reflection power versus depth were constructed at each study site. All radar traces within one Fresnel zone (3 km) of a central latitude and longitude were selected and averaged together to form a representative radar profile in a process similar to the incoherent summing

technique of Whitten et al. (2017). Before averaging, traces (vertical radar profiles) at each site were shifted to align their first returns and thus reduce any depth differences introduced by local surface slopes or relative shifts due to variable ionospheric delays in time. Polar night observations have little ionospheric delays, but each daytime observation has to be adjusted individually. The number of traces selected at each site varied between 16 and 184, with sites closer to the pole generally including a larger number of traces due to more dense radar coverage.

Once profiles were constructed, reflectors within 500 m of the surface were automatically chosen by selecting every point in the averaged profile that corresponded to a local maximum. This depth limit was chosen to constrain our analysis to the top reflector packet identified by Phillips et al. (2008) and Putzig et al. (2009), enabling future direct comparisons to previously mapped outcrop stratigraphy and limiting the effect of propagation losses. Depths were calculated from the radar time delay assuming a wave velocity below the surface consistent with propagation in pure ice ($\epsilon = 3.15$). While we assume some layers have dielectric constants greater than 3.15, such layers should be thin relative to the total depth, and thus, we assume they do not have a large effect on the average dielectric constant.

Observed reflection powers and calculated reflectivities are quoted here on a logarithmic decibel scale, where more negative numbers correspond to lower power and weaker reflectivity. Power values were taken directly from uncalibrated radargrams available on National Aeronautics and Space Administration's Planetary Data System, which were produced

Figure 4. Map of the cap-wide reflector used as context for this study. White dots denote sites for which radar profiles were created. Color corresponds to reflectivity measured using the technique described in section 3.2. Reflectivity is reported using the logarithmic decibel scale, where more negative numbers represent weaker reflectivities.

Figure 5. Reflection power profiles extracted and averaged from all traces at each site. Red dots correspond to local maxima selected as reflectors. Dotted lines mark the cap-wide reflector discussed in the text and shown in Figures 3 and 4. Each panel corresponds to a separate study site as labeled in Figure 4.

using the U.S. science team processor. In order to avoid erroneously identifying peaks in the background noise as reflectors, only reflectors with peak power above -35 dB were selected for analysis. This lower bound introduces the possibility that some particularly dim reflectors are ignored by the autodetection procedure. Such dim reflectors are likely rare within the first 500 m of the SHARAD record, given the low-loss nature of the NPLD (Grima et al., 2009) and the range of reflector powers previously observed (Lalich & Holt, 2016; Phillips et al., 2008). The auto-detection procedure might also fail to disqualify processing artifacts known as “range side lobes,” which may appear as faint reflections directly beneath the surface, though qualitative analysis of reflection profiles indicates that this is likely not a widespread issue. Final averaged profiles and selected reflections are shown in Figure 5.

These subsurface reflector powers are then used to calculate the reflectivity (i.e., the ratio of reflected power to incoming power) for each subsurface interface. This was done following the method from Lalich and Holt (2016), which is itself adapted from Lauro et al. (2012). In this approach, subsurface reflectivity (R_{ss}) is expressed as a function of observed reflection power at the surface and subsurface (respectively, P_s and P_{ss}), the surface reflectivity (R_s), and an estimate of power lost due to propagation in the medium. Our approach differs from that of previous work in two significant ways. First, instead of using a constant value for R_s derived from the assumed composition of the surface layer, we use empirically determined values for each site from Grima et al. (2012), selecting the surface reflectivity value from the location closest to each individual radar profile. This allows us to more accurately account for the impact of local surface properties on the calculation of subsurface reflectivity. Details regarding the assumptions inherent in these empirical values are available in Grima et al. (2012); we did not attempt to further quantify the uncertainty introduced by the use of these measurements.

The second significant change is the adoption of an empirically derived NPLD loss tangent of $\tan \delta = 0.0026$ (Grima et al., 2009) in order to express transmission losses, rather than using a theoretical value (Lauro et al.,

Figure 6. Schematic showing the various measured and modeled parameters discussed in section 3.

2012) or simply assuming negligible losses (Lalich & Holt, 2016). Like surface reflectivity, loss tangent also varies over the polar cap. However, measurements of loss tangent are not available outside of Gemina Lingula (Grima et al., 2009), so here we limit our analysis to the average value observed there. The uncertainties in the final dust content estimates introduced by this assumption are small for all but the deepest reflectors. Using loss tangent values one standard deviation away from the average introduced a relative error on the order of 10% at 300-m depth. Given the low value of the loss tangent, we adopt the approximation $\tan \delta \approx \delta$. This alteration results in the following expression to determine subsurface reflectivity:

$$R_{ss} = \frac{P_{ss}}{P_s} \frac{R_s e^{2\delta k z}}{(1 - R_s)^2}, \quad (1)$$

where k is the wavenumber at SHARAD's center frequency of 20 MHz and z is the depth below the surface. Reflectivity estimates were made for every identified reflector.

One other assumption made in accordance with previous work is that the roughness of each interface is the same (Lalich & Holt, 2016; Lauro et al., 2012). Previous work has shown that surface scattering due to roughness is low on the present-day surface (Grima et al., 2012), and we assume that this low roughness holds for at least the upper reflector packet of the NPLD.

3.3. Modeling Marker Bed Reflectivity and Estimating Dust Content

For this analysis, we assume that radar reflections observed by SHARAD are caused by marker beds, which are commonly believed to be thin, dust-rich layers (Christian et al., 2013; Fishbaugh et al., 2010; Lalich & Holt, 2016). Marker beds are typically thinner than SHARAD's range resolution, meaning they cannot be modeled as a simple interface. Instead, we apply the technique developed by MacGregor et al. (2011) to model the effective reflectivity of two interfaces that are closely spaced relative to radar wavelengths, that is, the top and bottom of a thin layer. This model was previously used to simulate marker bed reflectivity by Lalich and Holt (2016) and can be expressed as follows:

$$R = \left| r_{01} + t_{10} r_{12} t_{01} \left(\frac{e^{-2ik\Delta}}{1 - r_{12} r_{10} e^{-2ik\Delta}} \right) \right|^2, \quad (2)$$

where reflectivity (R) is expressed in terms of the Fresnel reflection and transmission coefficients at each interface (r_{xx} and t_{xx} where the first subscript is the layer through which the wave is propagating and the second is the layer it is incident upon), the complex wave number in the thin layer (k), and the thickness (Δ) of that layer (see Figure 6). In order to calculate the Fresnel reflection and transmission coefficients, we first assume the thin layer is composed of a dust and ice mixture and is bounded above and below by pure water ice. Then, we calculate a range of permittivities for this mixture using the Maxwell-Garnett mixing formula (Garnett, 1906), with pure ice and pure dust as end members. This necessitates selecting a permittivity value for the dust inclusions. Nunes and Phillips (2006) proposed three potential values for Martian dust permittivity (to be mixed with the value of 3.15 for ice). We selected the intermediate value ($\epsilon = 8.8 - 0.017i$) for this work, which is the value corresponding to a shergottite. Using this range of mixture permittivities, we are able to calculate the reflection and transmission coefficients necessary for the model input for each dust content value. In order to construct a parameter space of possible layer configurations, we consider a range of thicknesses between 1 and 10 m. We then use equation (2) to calculate the reflectivity for each dust content-thickness pair.

In contrast to Lalich and Holt (2016), who limited the model to SHARAD's center frequency, here we include the full bandwidth of the transmitted chirp signal. By repeating the procedure outlined above for every frequency sampled by the SHARAD chirp, we can build a three dimensional parameter space of reflectivity as a function of dust content, layer thickness, and frequency. As SHARAD is limited to a finite bandwidth, the

Figure 7. Modeled thin layer reflectivity for bed thicknesses between 0 and 10 m and fractional dust contents between 1% and 100%. Layers less than 1 m thick (indicated by dark area above) were not considered during dust content estimation.

Fourier transform of the transmitted chirp can be interpreted as an energy density function, describing how much relative power is transmitted at each frequency. SHARAD does not transmit power evenly across its bandwidth (Campbell & Morgan, 2018), so rather than using a simple linear chirp, we use the SHARAD reference chirp obtained from National Aeronautics and Space Administration's Planetary Data System, which was determined from ground tests.

Multiplying our parameter space across the frequency axis by this density function and then integrating over all frequencies allows us to express the total reflectivity as observed by SHARAD as a function of only dust content and layer thickness while accounting for the nonuniform distribution of energy across the frequency bandwidth (see Figure 7).

This new model differs significantly from those shown by Lalic and Holt (2016) and allows for more direct comparisons with radar observations. Importantly, contours of constant reflectivity are limited to narrower ranges of dust content. For most measured reflectivities, dust content varies by approximately 20%, though the inferred dust content of reflectors with high reflectivity can vary more due to the increased sensitivity of the model between 2- and 4-m thicknesses. This means that by measuring SHARAD reflectivity as outlined above, it is possible to place constraints on the fractional dust content of individual layers. For every layer thickness value, we record the dust content needed for the modeled reflectivity to match a given observed reflectivity. Assuming all modeled layer thicknesses from 1- to 10-m thick are equally probable for a given reflector, this allows us to construct a distribution of possible dust content values for each analyzed reflector. An example of this procedure is shown in Figure 8. The results of this procedure are sensitive to the thickness range selected for the model, which could bias the final dust content distribution to low or high values. The 1- to 10-m range used here was selected based on measurements of marker bed thickness and SHARAD's nominal resolution in water ice.

4. Results and Discussion

4.1. Reflectivity as a Dust Proxy

When the above procedure is applied to the study sites discussed in section 3.1, we retrieve a wide range of possible dust contents. There were between 13 and 19 reflectors identified at each site within the top 500 m of the NPLD. Plots of the median inferred dust content for each reflector at each site as a function of depth are shown in Figure 9. These results show that reflectors can vary between ~10% and 60% dust vertically within a single site, implying drastic differences in surface conditions through time.

Figure 8. Outline of process for estimating dust content. (a) Reflection power versus depth for site B. Black box shows which reflector is used for (b) and (c). Annotations show values for this reflector's measured reflection power (P_{ss}) and reflectivity (R_{ss}). (b) The model described in section 3.3, with a contour of constant reflectivity at the value shown in (a) or -16.5 dB (indicated by black line on color scale bar). (c) The distribution of all dust content values for the contour shown in (b), along with the median and standard deviation. We use the median rather than the mean in order to more accurately capture the center of the distribution, which features a long tail caused by layer thicknesses between ~ 2 and 4 m. In this context, the standard deviation can be viewed as an error estimate, influenced in part by the strength of this tail.

We have constrained the dust content of individual reflectors to a relatively narrow range, but there is still noticeable uncertainty in this estimation due to the ambiguity in layer thickness. As shown in Figures 7 and 8, there is an excursion in modeled reflectivity toward high dust content values for layer thicknesses between roughly 2 and 4 m. This excursion results in a long tail in each reflector's modeled dust distribution, which leads to an alternate interpretation of reflectivity differences between reflectors. Instead of these reflectivity differences being the result of large changes in dust content, they could instead be smaller fluctuations in layer thickness. This hypothesis was examined in more detail by Lalich and Holt (2016). They concluded that there was superficial correlation between regional radar reflectivity and reflector depth, implying that areas with higher or lower average accumulation rates contained consistently thicker and thinner dust-rich layers, respectively, and this was manifesting in the radar record as variable subsurface reflectivity.

According to the improved full-bandwidth model used here, most layers would have to be ~ 2 – 4 m thick, the range where reflectivity is most sensitive to the separation of the layer interfaces, in order for thickness variation to explain reflectivity differences. However, other studies have identified marker beds with thicknesses well in excess of 4 m, more than half the resolution of SHARAD in ice (e.g., Becerra et al., 2016; Fishbaugh et al., 2010). According to our model, small thickness variations in such layers would not produce notable changes in reflectivity. Becerra et al. (2016) also searched for a correlation between average deposition rates and marker bed thickness and found no statistically significant relationship. Together, these observations favor the hypothesis that the primary cause of reflectivity variation for most layers is in fact differences in dust content and not layer thickness. However, individual layers may be within the 2- to 4-m thickness range, where reflectivity depends more strongly on thickness than dust content.

Another interesting result concerns the possible presence and detectability of pure dust lag deposits. The creation of a dust lag is a self-limiting process. Once the lag becomes thick enough, it protects the ice

Figure 9. Profiles of median fractional dust content versus depth for each analyzed profile. Vertical error bars are ± 4.5 m, in accordance with Shallow Radar's theoretical resolution in water ice but not taking into account vertical smearing from side lobe suppression. Horizontal error bars are the standard deviation of the dust distribution for each reflector. Dotted lines mark the cap-wide reflector discussed in the text and shown in Figures 3 and 4. Each panel corresponds to a single study site as labeled in Figure 4.

below it from further sublimation, and thus, the lag itself stops growing in thickness (Bramson et al., 2017; Bramson et al., 2019). Previous work has found this limit to be as thin as a few millimeters (Skorov et al., 2001) or as thick as ~ 5 cm (Paige, 1992; Paige et al., 1994) for the polar regions of Mars. According to our reflectivity model, a pure dust layer with a reflectivity of -21.8 dB (the median across all study sites, see Table 1) would need to be at least 11 cm thick, more than double the maximum thickness of a dust lag predicted by the previous studies. Though they may exist, any pure dust layers thinner than ~ 5 cm would not produce bright enough reflections to be detected above SHARAD's noise floor. Therefore, we interpret all bright reflections to be evidence of thick dust rich layers, not dust lags, supporting the marker bed-reflector assumption.

One potential limitation of this analysis is that the modeled layer geometry may be overly simplistic. For example, rather than reflectors being caused by a single layer, they could in fact be the result of multiple thin layers within a single range resolution cell (Campbell & Morgan, 2018). While such geometries are possible, there is little observational evidence connecting them to radar reflectors at this time, in contrast to the genetic link drawn between SHARAD reflectors and marker beds (Christian et al., 2013). Therefore, we interpret the present evidence to conclude that multiple thin layers are not the dominant mechanism for causing bright SHARAD reflections.

4.2. Site Variability

Mean dust content of the several dusty layers at each site varies from 17% to 36% or roughly varying by a factor of 2. However, if most layers are in the 2- to 4-m range discussed previously, layer dust contents could be much higher. The mean dust contents reported in Table 1 are

Table 1
Comparison of Mean Dust Content for Dusty Layers and Cap-Wide Reflector Dust Content at Each Study Site

Dust content	Site									
	A	B	C	D	E	F	G	H	I	J
Cap-wide reflector dust content (%)	30	40	20	16	35	21	18	17	17	34
Mean dust content (%)	17	36	28	27	28	17	17	19	23	27

dominated by the model region outside of this range. Because the bulk dust content of the NPLD was determined to be only 5% (Grima et al., 2009), thick, dust-rich layers could increase the profile average above a total of 5% for the uppermost 500 m, in potential disagreement with those results. For example, if most layers are near 10 m thick, then bulk dust content at most sites exceeds the 5% average set by Grima et al. (2009); however, if dusty layers tend to be near 5 m thick, then most sites fall beneath this threshold. This could be viewed as either a constraint on dusty layer thickness or an indication that we are systematically overestimating dust content.

Large variations can be seen within each individual site. These variations show the importance of constraining small-scale regional differences in ice and dust deposition in future work and suggest a cautionary approach when inferring global climate information from NPLD stratigraphy. In addition, as shown in Table 1, there is no simple correlation between the dust content of the cap wide reflector and the mean dust content at each study site. Instead, it appears that the surface distribution of dust on the NPLD changes with time.

It is notable that the number of reflectors identified at each site is not equal. The cap-wide horizon discussed here represents an unconformity at many locations, with corresponding loss of icy and dusty material (Smith et al., 2016). The presence of such a widespread unconformity indicates it is probable that many reflectors at each site are not laterally continuous across the entire cap and do not appear in the record sampled by every profile. Reflectors could also be interfering with each other or be too closely spaced, causing some to go unresolved. This hypothesis is supported by the fact that the profiles with the most reflectors are at sites where the top layer packet is the thickest, indicating that deposition rates were generally higher in these areas during packet accumulation, which causes layers to be spaced further apart. The mean separation distance between reflectors is 25 ± 11 m (using a bulk dielectric content of 3.15 to estimate depth), which is near the estimated reflector separation resolution of ~ 20 m for SHARAD in the NPLD (Nunes & Phillips, 2006). Indeed, reflectors connecting the three study sites in the saddle region often merge together and then split again, making it difficult to definitively trace reflectors between sites even when they are not separated by troughs or scarps. Future work aiming to use radar stratigraphy as a climate record should be cognizant of this effect and conducted where the cap is thickest, in order to ensure that the stratigraphic record is as complete as possible.

Taken together, these observations indicate that the accumulation of ice and dust in the NPLD is greatly influenced by regional conditions such as wind speed and direction or even stochastic events like impacts, which may vary over different time and distance scales than the orbital forcing typically invoked to explain layer formation.

4.3. Layer Formation Processes and Climate Model Comparisons

As discussed in section 2.3, previous work found that reflector-causing layers are formed through two main processes, both of which are primarily controlled by the planet's orbital parameters: the concentration of dust through ice sublimation (Levrard et al., 2007; Toon et al., 1980) or the nonuniform variation of ice and dust deposition with time (e.g., Cutts & Lewis, 1982; Hvidberg et al., 2012; Phillips et al., 2008; Putzig et al., 2009). Each process produces distinct populations of layers. Lag deposits consist almost entirely of dust and either empty or ice-filled pore space, while constructional ice layers may have much lower fractional dust contents, between 3% and 10% according to a recent model of obliquity-driven polar ice and dust deposition (Hvidberg et al., 2012). However, our dust content estimates do not recreate this bimodal behavior. A full distribution of dust content for analyzed layers is shown in Figure 10. We find only two reflectors for which dust content is less than 10% and only four of the 156 analyzed reflectors have a dust content over 50%. The dearth of reflections caused by nearly pure dust is no surprise, given that lags thicker than a few centimeters cannot form. Not only do we fail to observe the predicted bimodal distribution, we also find that for our chosen reflectors, dust contents do not match those predicted for either formation process. However, there are some caveats to our analysis. First, reflectors that are less than -35 dB would represent cases with very low dust concentrations, and we do not attempt to separate any such reflectors from the surrounding noise. Second, we only consider 10 locations across a very large ice cap, and the missing bimodal distribution may be lost in small number statistics. Lastly, as previously discussed, our choice of modeled layer thicknesses could bias our final dust content estimates.

Figure 10. Histogram of median estimated dust fraction for each reflector across all sites. Bin size is 2%.

One potential explanation for the lack of a bimodal dust content distribution is that bright reflectors caused by high dust content layers are masking or blending with dimmer reflectors caused by low dust content layers that might otherwise be present in the radar record. Bright reflectors are identified with a spacing close to SHARAD's resolution in the upper reflector packet. Any layers with a fractional dust content below 10% that might be between these higher dust content layers simply would not be resolved by SHARAD.

It is also possible that low dust content layers exist in the interpacket regions between the main groups of SHARAD reflectors (see Figure 2), but they may not cause strong enough reflections to be visible above our chosen -35 -dB threshold. According to our reflectivity model, a 6-m-thick layer with 5% fractional dust content would have a reflectivity of -34.68 dB. At 400 m below the surface (approximately the bottom of the first reflector packet), this would correspond to an observed reflection power of -46.73 dB, which is below SHARAD's nominal noise floor of -40 dB. This is consistent with the accumulation model of Hvidberg et al. (2012), where low dust content layers are predicted throughout deposition, and the bright packet structure that defines SHARAD stratigraphy in the NPLD is interpreted as the result of layers formed during periods of abnormally low ice deposition. These low ice deposition layers would

only form during periods that contained swings of maximum obliquity, confining them to specific times in the stratigraphic record, similar to the observed reflector packets (Phillips et al., 2008; Putzig et al., 2009).

Given the apparent lack of lag deposits, we require another explanation for the relatively high ($>10\%$) dust contents observed in the NPLD. Hvidberg et al. (2012) modeled both dust and ice deposition as anticorrelated with obliquity, that is, higher obliquity resulted in lower deposition rates of ice and dust and vice versa. However, polar dust deposition rates may not be so simply correlated with obliquity. Newman et al. (2005) modeled the dust cycle of Mars under different orbital conditions and found that dust deposition rates at the poles increase as obliquity departs from 25° , whether that departure is toward higher or lower obliquities. Thus, dust deposition rates would go through two full cycles of maximum and minimum deposition for each obliquity cycle. Notably, this means dust deposition rates would actually be approaching a maximum at the same time the modeled ice deposition rates of Hvidberg et al. (2012) are approaching their minimum during periods of high obliquity. Their model placed minimum ice deposition rates on the order of ~ 0.1 mm/year and maximum dust deposition rates near ~ 0.02 mm/year, which when combined would result in a fractional dust content of $\sim 17\%$, similar to the values estimated here. In order to determine if this covarying behavior of ice and dust deposition is a plausible mechanism for creating reflector-causing layers, future studies should focus on how the global dust cycle changes over geologic time.

5. Conclusion

We presented a method for using SHARAD reflection power as a proxy for determining dust content in the NPLD. By measuring the reflectivity of an individual reflector and comparing it to a model, the range of possible dust contents for the reflector-causing layer can be constrained to a narrow distribution, though the impact of changing layer thickness should not be entirely discounted. Reflectivities greater than -10 dB are not found, implying that dust content never reaches 75% even for relatively thin layers. Pure dust lags should only reach a few centimeters thick and never generate a strong radar reflection at SHARAD wavelengths. Therefore, lags are not responsible for SHARAD reflections.

Mean reflectivity at each of our study sites does not correlate with the local reflectivity of a single cap-wide reflector, indicating changes in the geographic distribution of ice and dust deposition over time. In addition, the number of reflectors identified between sites is inconsistent, possibly as a result of erosional unconformities, nonuniform deposition causing layers to pinch out, or reflectors becoming too weak or too close together to be separately resolved by SHARAD. Together, these two observations indicate that regional or

local processes play a large role in accumulation. This must be carefully accounted for if the NPLD is to be used as a global climate record from orbital-based observations.

Estimates of fractional dust content do not display the bimodal distribution predicted by some models. Instead, it is possible that low dust content layers that might otherwise cause reflections in the upper reflector packet are simply masked by much brighter high dust reflectors. While low dust content layers might be present in the interpacket regions of the NPLD, any reflections they cause are likely below SHARAD's noise floor. The presence of nearly pure dust lag deposits cannot be confirmed from our calculations, as they would be too thin, and indeed, we find it unlikely that SHARAD could detect such a thin layer. We show that the intermediate dust concentrations estimated here could instead be explained by altering a previously published NPLD accumulation model (Hvidberg et al., 2012) to account for periods of sustained accumulation with higher than average dust fraction.

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References

- Becerra, P., Byrne, S., Sori, M. M., Sutton, S., & Herkenhoff, K. E. (2016). Stratigraphy of the north polar layered deposits of Mars from high-resolution topography. *Journal of Geophysical Research: Planets*, 121, 1445–1471. <https://doi.org/10.1002/2015JE004992>
- Becerra, P., Sori, M. M., & Byrne, S. (2017). Signals of astronomical climate forcing in the exposure topography of the north polar layered deposits of Mars. *Geophysical Research Letters*, 44, 62–70. <https://doi.org/10.1002/2016GL071197>
- Bramson, A. M., Byrne, S., & Bapst, J. (2017). Preservation of midlatitude ice sheets on Mars. *Journal of Geophysical Research: Planets*, 122, 2250–2266. <https://doi.org/10.1002/2017JE005357>
- Bramson, A. M., Byrne, S., Bapst, J., Smith, I. B., & McClintock, T. (2019). A migration model for the polar spiral troughs of Mars. *Journal of Geophysical Research: Planets*, 124, 1020–1043. <https://doi.org/10.1029/2018JE005806>
- Brown, A. J., Calvin, W. M., Becerra, P., & Byrne, S. (2016). Martian north polar cap summer water cycle. *Icarus*, 277, 401–415. <https://doi.org/10.1016/j.icarus.2016.05.007>
- Campbell, B. A., & Morgan, G. A. (2018). Fine-scale layering of Mars polar deposits and signatures of ice content in nonpolar material from multiband SHARAD data processing. *Geophysical Research Letters*, 45, 1759–1766. <https://doi.org/10.1002/2017GL075844>
- Christian, S., Holt, J. W., Byrne, S., & Fishbaugh, K. E. (2013). Integrating radar stratigraphy with high resolution visible stratigraphy of the north polar layered deposits, Mars. *Icarus*, 226(2), 1241–1251. <https://doi.org/10.1016/j.icarus.2013.07.003>
- Cutts, J. A., & Lewis, B. H. (1982). Models of climate cycles recorded in Martian polar layered deposits. *Icarus*, 50(2–3), 216–244. [https://doi.org/10.1016/0019-1035\(82\)90124-5](https://doi.org/10.1016/0019-1035(82)90124-5)
- Fishbaugh, K. E., Byrne, S., Herkenhoff, K. E., Kirk, R. L., Fortezzo, C., Russell, P. S., & McEwen, A. (2010). Evaluating the meaning of “layer” in the Martian north polar layered deposits and the impact on the climate connection. *Icarus*, 205(1), 269–282. <https://doi.org/10.1016/j.icarus.2009.04.011>
- Fishbaugh, K. E., & Hvidberg, C. S. (2006). Martian north polar layered deposits stratigraphy: Implications for accumulation rates and flow. *Journal of Geophysical Research*, 111, E06012. <https://doi.org/10.1029/2005JE002571>
- Garnett, J. C. M. (1906). Colours in metal glasses, in metallic films, and in metallic solutions. II. *Philosophical Transactions of the Royal Society of London. Series A, Containing Papers of a Mathematical or Physical Character*, 205, 237–288.
- Greve, R., Grieger, B., & Stenzel, O. J. (2010). MAIC-2, a latitudinal model for the Martian surface temperature, atmospheric water transport and surface glaciation. *Planetary and Space Science*, 58(6), 931–940. <https://doi.org/10.1016/j.pss.2010.03.002>
- Grima, C., Kofman, W., Herique, A., Orosei, R., & Seu, R. (2012). Quantitative analysis of Mars surface radar reflectivity at 20 MHz. *Icarus*, 220(1), 84–99. <https://doi.org/10.1016/j.icarus.2012.04.017>
- Grima, C., Kofman, W., Mougnot, J., Phillips, R. J., Hérique, A., Biccari, D., et al. (2009). North polar deposits of Mars: Extreme purity of the water ice. *Geophysical Research Letters*, 36, L03203. <https://doi.org/10.1029/2008GL036326>
- Herkenhoff, K. E., Byrne, S., Russell, P. S., Fishbaugh, K. E., & McEwen, A. S. (2007). Meter-scale morphology of the north polar region of Mars. *Science*, 317(5845), 1711–1715. <https://doi.org/10.1126/science.1143544>
- Herkenhoff, K. E., & Plaut, J. J. (2000). Surface ages and resurfacing rates of the polar layered deposits on Mars. *Icarus*, 144(2), 243–253. <https://doi.org/10.1006/icar.1999.6287>
- Hvidberg, C. S., Fishbaugh, K. E., Winstrup, M., Svensson, A., Byrne, S., & Herkenhoff, K. E. (2012). Reading the climate record of the Martian polar layered deposits. *Icarus*, 221(1), 405–419. <https://doi.org/10.1016/j.icarus.2012.08.009>
- Lalich, D. E., & Holt, J. W. (2016). New Martian climate constraints from radar reflectivity within the north polar layered deposits. *Geophysical Research Letters*, 44, 657–664. <https://doi.org/10.1002/2016GL071323>
- Landis, M. E., Byrne, S., Daubar, I. J., Herkenhoff, K. E., & Dundas, C. M. (2016). A revised surface age for the north polar layered deposits of Mars. *Geophysical Research Letters*, 43, 3060–3068. <https://doi.org/10.1002/2016GL068434>
- Laskar, J., Levrard, B., & Mustard, J. F. (2002). Orbital forcing of the Martian polar layered deposits. *Nature*, 419(6905), 375–377. <https://doi.org/10.1038/nature01066>
- Lauro, S. E., Mattei, E., Soldovieri, F., Pettinelli, E., Orosei, R., & Vannaroni, G. (2012). Dielectric constant estimation of the uppermost Basal Unit layer in the Martian Boreales Scopuli region. *Icarus*, 219(1), 458–467. <https://doi.org/10.1016/j.icarus.2012.03.011>
- Levrard, B., Forget, F., Montmessin, F., & Laskar, J. (2007). Recent formation and evolution of northern Martian polar layered deposits as inferred from a Global Climate Model. *Journal of Geophysical Research*, 112, E06012. <https://doi.org/10.1029/2006JE002772>
- MacGregor, J. A., Anandakrishnan, S., Catania, G. A., & Winebrenner, D. P. (2011). The grounding zone of the Ross Ice Shelf, West Antarctica, from ice-penetrating radar. *Journal of Glaciology*, 57(205), 917–928. <https://doi.org/10.3189/002214311798043780>
- Malin, M. C., & Edgett, K. S. (2001). Mars Global Surveyor Mars Orbiter Camera: Interplanetary cruise through primary mission. *Journal of Geophysical Research*, 106(E10), 23,429–23,570. <https://doi.org/10.1029/2000JE001455>
- McEwen, A. S., Soderblom, L. A., Becker, T. L., Lee, E. M., Swann, J. D., Aeschliman, R., & Batson, R. M. (1994). Global color views of Mars. (Vol. 25, p. 871). Presented at the Lunar and Planetary Science Conference
- Milkovich, S. M., & Head, J. W. (2005). North polar cap of Mars: Polar layered deposit characterization and identification of a fundamental climate signal. *Journal of Geophysical Research*, 110, E01005. <https://doi.org/10.1029/2004JE002349>

- Milkovich, S. M., Head, J. W. III, & Neukum, G. (2008). Stratigraphic analysis of the northern polar layered deposits of Mars: Implications for recent climate history. *Planetary and Space Science*, 56(2), 266–288. <https://doi.org/10.1016/j.pss.2007.08.004>
- Newman, C. E., Lewis, S. R., & Read, P. L. (2005). The atmospheric circulation and dust activity in different orbital epochs on Mars. *Icarus*, 174(1), 135–160. <https://doi.org/10.1016/j.icarus.2004.10.023>
- Nunes, D. C., & Phillips, R. J. (2006). Radar subsurface mapping of the polar layered deposits on Mars. *Journal of Geophysical Research*, 111, E06S21. <https://doi.org/10.1029/2005JE002609>
- Paige, D. A. (1992). The thermal stability of near-surface ground ice on Mars. *Nature*, 356(6364), 43–45. <https://doi.org/10.1038/356043a0>
- Paige, D. A., Bachman, J. E., & Keegan, K. D. (1994). Thermal and albedo mapping of the polar regions of Mars using Viking thermal mapper observations: 1. North polar region. *Journal of Geophysical Research*, 99(E12), 25,959–25,991. <https://doi.org/10.1029/93JE03428>
- Perron, J. T., & Huybers, P. (2009). Is there an orbital signal in the polar layered deposits on Mars? *Geology*, 37(2), 155–158. <https://doi.org/10.1130/G25143A.1>
- Phillips, R. J., Zuber, M. T., Smrekar, S. E., Mellon, M. T., Head, J. W., Tanaka, K. L., et al. (2008). Mars north polar deposits: Stratigraphy, age, and geodynamical response. *Science*, 320(5880), 1182–1185. <https://doi.org/10.1126/science.1157546>
- Putzig, N. E., Phillips, R. J., Campbell, B. A., Holt, J. W., Plaut, J. J., Carter, L. M., et al. (2009). Subsurface structure of Planum Boreum from Mars Reconnaissance Orbiter Shallow Radar soundings. *Icarus*, 204(2), 443–457. <https://doi.org/10.1016/j.icarus.2009.07.034>
- Seu, R., Phillips, R. J., Biccari, D., Orosei, R., Masdea, A., Picardi, G., et al. (2007). SHARAD sounding radar on the Mars Reconnaissance Orbiter. *Journal of Geophysical Research*, 112, E05S05. <https://doi.org/10.1029/2006JE002745>
- Skorov, Y. V., Markiewicz, W. J., Basilevsky, A. T., & Keller, H. U. (2001). Stability of water ice under a porous nonvolatile layer: Implications to the south polar layered deposits of Mars. *Planetary and Space Science*, 49(1), 59–63. [https://doi.org/10.1016/S0032-0633\(00\)00121-5](https://doi.org/10.1016/S0032-0633(00)00121-5)
- Smith, I. B., & Holt, J. W. (2015). Spiral trough diversity on the north pole of Mars, as seen by Shallow Radar (SHARAD). *Journal of Geophysical Research: Planets*, 120, 362–387. <https://doi.org/10.1002/2014JE004720>
- Smith, I. B., Putzig, N. E., Holt, J. W., & Phillips, R. J. (2016). An ice age recorded in the polar deposits of Mars. *Science*, 352(6289), 1075–1078. <https://doi.org/10.1126/science.aad6968>
- Tanaka, K. L. (2005). Geology and insolation-driven climatic history of Amazonian north polar materials on Mars. *Nature*, 437(7061), 991–994. <https://doi.org/10.1038/nature04065>
- Tanaka, K. L., Rodriguez, J. A. P., Skinner, J. A., Bourke, M. C., Fortezzo, C. M., Herkenhoff, K. E., et al. (2008). North polar region of Mars: Advances in stratigraphy, structure, and erosional modification. *Icarus*, 196, 318–358. <https://doi.org/10.1016/j.icarus.2008.01.021>
- Toon, O. B., Pollack, J. B., Ward, W., Burns, J. A., & Bilski, K. (1980). The astronomical theory of climatic change on Mars. *Icarus*, 44(3), 552–607. [https://doi.org/10.1016/0019-1035\(80\)90130-X](https://doi.org/10.1016/0019-1035(80)90130-X)
- Whitten, J. L., Campbell, B. A., & Morgan, G. A. (2017). A subsurface depocenter in the south polar layered deposits of Mars. *Geophysical Research Letters*, 44, 8188–8195. <https://doi.org/10.1002/2017GL074069>