

**Detangling the processes recorded in ice: Geomorphic and hyperspectral analysis of
Mars' north polar spiral troughs**

By

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A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of
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Thesis: "The formation of irregular polygonal ridge networks, Nili Fossae, Mars: Implications for extensive subsurface channelized fluid flow in the Noachian"

Advisor: John F. Mustard

Smith College, Northampton, MA (Sept 2011 – May 2015)

B. A., Geoscience and Astronomy, Departmental High Honors

Senior Honors Thesis: "Geologic Mapping and Characterization of Nicholson Crater, Mars"

Advisor: Robert Newton



Notable Datasets

Mars Recon. Orbiter: CRISM, HiRISE, CTX | Mars Express: HRSC, OMEGA | Mars Global Surveyor: MOLA | Mars Odyssey: THEMIS

Topics of Expertise

VNIR-IR Spectroscopy | Geology (Geomorphology, Mineralogy, Sedimentology / Stratigraphy, Surface Processes) | Remote Sensing | Radiative Transfer Theory (Hapke and Shkuratov)

Computer Languages

Highly Proficient in Matlab | Experience in IDL and Python | Operating system fluency in Windows and Mac with experience in UNIX

Software

ENVI | ArcMap and ArcScene | JMARS | ISIS3/SPICE | Craterstats2 | Adobe Suite: Illustrator, Photoshop, InDesign, PremierPro

RESEARCH EXPERIENCE

Graduate Student Researcher/Ph.D. Student, Brown University, 2015-present

Data Processing and Computer Programming Skills

- Highly experienced in image analysis and data processing of imagery, multispectral, and hyperspectral data. Processing remotely sensed data from raw DN to reflectance (or I/F), and further processing of optical data, including reflectance to single-scattering albedo.
- Highly experienced in and deep understanding of various types/degrees of photometric and atmospheric corrections for remotely sensed data.
- Experience using UNIX and ISIS3/SPICE to process HiRISE data. Bash script writing and HiRISE DTM production.
- Quantitative and qualitative spectral modeling in both ENVI and Matlab (custom codes). Specific expertise in quantitative spectral mixture modeling of ices/ice mixtures.
- Highly experienced with processing a variety of optical planetary datasets. Especially with NASA Planetary Data System data retrieval.
- High comfort level in learning to troubleshoot, adapt, and use new datasets.
- Skilled in identifying, categorizing, classifying, and interpreting geologic features and processes in ArcMap. Production of feature class databases and maps for sub-regions of Mars.
- Experience in crater counting in ArcMap and crater age dating of planetary surfaces using Craterstats2 software.

Research Accomplishments

- Developed a hyperspectral processing workflow to investigate the influence of atmospheric aerosols and surface scattering behavior on VNIR-SWIR data, expertise with CRISM.
- Development and use of linear and non-linear spectral modeling and (forward and inverse) radiative transfer models to estimate grain size and abundance of planetary surface materials (i.e., minerals) from spectral data.
- Geo-referencing multiple overlapping datasets for data integration and analysis ease of use (e.g., HiRISE-CTX-CRISM geo-referencing).

Undergraduate Researcher, Advisor: Dr. Robert Newton, Smith College, AY 2014

- Geomorphic mapping of aeolian and periglacial features on Mars, specifically in Nicholson Crater (Context Camera and HiRISE imagery)
- Determined the timing and sequence of geologic events.

Lunar & Planetary Science Institute Undergraduate Research Intern,

Mentor: Dr. Carlton Allen, Summer 2014

- Survey of martian craters using the Mars Crater Database for craters with large enigmatic sedimentary deposits.
- Investigated possible processes responsible for forming large sedimentary deposits located in the center of 6/255 surveyed craters through geomorphic analysis. Proposed primary deposition and erosion mechanisms.

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Honors & Recognitions

- JPL Planetary Science Summer School Alum, Proposal Manager of TIGER (PSSS.S1 2020 conceptmission to Enceladus)
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- Smith College Dean's List, 2012-2015
- Elected for SigmaXi, 2015
- Academic All-Conference Honors: Scholar-Athlete >3.5 GPA, Swimming – NEWMAC, 2015
- Featured in Smith College Alumnae Relations Magazine, One Student: To this geoscience major, Mars rocks by Christina Barber-Just, Winter 2014.

PROFESSIONAL SERVICES AND CONTRIBUTIONS

- NASA Executive Secretary, Review Panel, 2020
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Co-advised with John F. Mustard.

- Mentored sophomore undergraduate through her first, independent, planetary research project titled, "North Polar Layered Deposits of Mars: Investigating the effects of perennial dust deposits on water ice sublimation."
- Taught mentee how to acquire planetary data of Mars from the PDS, use ArcMap, and code in Matlab.
- Held mentor-mentee meetings where I engaged my mentee in exercises to develop scientific articles critiquing skills, and developing and approaching a scientific question.

DEEP Science Teaching and Education Program (DEEP STEP) Co-Coordinator, 2017 – 2021

- Developed quarter-long, multi-lesson Earth Science Modules for grades 2-5 for the Providence public school district (aligned with the Next Generation Science Standards) centered on inquiry- driven science learning.
- Lead the development of modules for 2nd and 4th grade.
- Co-taught the lessons with elementary school teachers with the goal of developing their confidence in teaching science independently.

Brown University Teaching

Lead Course Designer

GEOL 0050: Mars, Moon and the Earth (undergraduate non-major) (2018-2019)

Teaching Assistant

GEOL 0050: Mars, Moon and the Earth (undergraduate non-major) (F2016; F2019)

GEOL 1330: Global Environmental Remote Sensing (undergraduate major) (S2018; S2019)

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GEOL 1330: Global Environmental Remote Sensing (undergraduate major) (S2018; S2019)

- Instructed a classroom of 10 students, 6 hours a week. Mentored students through individual end-of course research projects.
- Wrote a beginner's guide to ENVI for students.
- Developed exercises which used a diverse range of Earth satellite datasets such as AVIRIS, Landsat, and IKONOS. Exercises tapped into various image analysis techniques and products such as NDVI, NBR, NDWI, (Un)Supervised Classification, SMA, SFF, and PCA.

Smith College Teaching

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Astronomy AST102: Sky I: Time (undergraduate non-major) Astronomy AST103: Sky II: Telescopes (undergraduate non-major)

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- Pascuzzo, A.C.,** Condu, T., Li, S., Mustard, J.F. (2021). Sensitivity analysis of ice/dust aerosol and surface phase function assumptions on Hapke models and band depth parameters over the martian north pole, *Icarus*, (*under review*)
- Tarnas, J.D., Mustard, J.F., Wu, X., Das, E., Cannon, K.M., Hundal, C.B., **Pascuzzo, A.C.,** Kellner, J.R., Parente, M., (2021). Successes and challenges of factor analysis/target transformation application to visible-to-near-infrared hyperspectral data, *Icarus*, 114402. <https://doi.org/10.1016/j.icarus.2021.114402>
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- Pascuzzo, A.C.,** Mustard, J.F., Kremer, C.H., Ebinger, E., (2018). The formation of irregular polygonal ridge networks, Nili Fossae, Mars: Implications for extensive subsurface channelized fluid flow in the Noachian. *Icarus* 319, 852-868. <https://doi.org/10.1016/j.icarus.2018.10.020>
- Johnson, B.C., Sheppard, R.Y., **Pascuzzo, A.C.,** Fisher, E.A., Wiggins, S.E., (2017). Porosity and salt content determine if subduction can occur in Europa's ice shell. *Journal of Geophysical Research: Planets* 122, 2765-2778. <https://doi.org/10.1002/2017JE005370>
- Deutsch, A.N., Head, J.W., Ramsley, K.R., Pieters, C.M., Potter, R.W.K., Palumbo, A.M., Bramble, M.S., Cassanelli, J.P., Jawin, E.R., Jozwiak, L.M., Kaplan, H.H., Lynch, C.F., **Pascuzzo, A.C.,** Qiao, L., Weiss, D.K., (2017). Science exploration architecture for Phobos and Deimos: The role of Phobos and Deimos in the future exploration of Mars. *Advances in Space Research, Past, Present and Future of Small Body Science and Exploration* 62, 2174-2186. <https://doi.org/10.1016/j.asr.2017.12.017>

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- Pascuzzo, A.C.,** T. Condu, J.F. Mustard, R.E. Arvidson, (2019). The effects of ice and dust aerosols, and surface scattering on the interpretation of the martian north polar ice surface characteristics using CRISM VNIR-SWIR data. 9th International Conference on Mars, Abstract #6380.
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- Pascuzzo, A.C.**, J.F. Mustard, (2018). The quest to derive the north polar layered deposits' compositional and physical properties through Hapke-radiative transfer modeling of CRISM hyperspectral data. Mars Workshop on Amazonian and Present-Day Climate, Abstract #4040.
- Pascuzzo, A.C.**, J.F. Mustard, E. Ebinger, (2018). The origin of enigmatic ridge networks, Nili Fossae, Mars: Implications for extensive subsurface fluid flow in the Noachian. LPSC 49, Abstract #2268.
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- *Mustard, J.F., and **A.C. Pascuzzo**, (2017). Late Noachian/Early Hesperian Ridge Networks in Nili Fossae: Evidence for Water-Saturated Near-Surface Crust, Abstract P24B-07 presented at 2017 Fall Meeting, AGU, New Orleans, LA, 11-15 Dec.
- Pascuzzo, A.C.**, B. C. Johnson, R.Y. Sheppard, E.A. Fisher, S.E. Wiggins, (2017). Porosity and salt content determine if subduction can occur in Europa's ice shell. Europa Deep Dive I: Ice-Shell Exchange Processes, Abstract #7006.
- ***Pascuzzo, A.C.** and J.F. Mustard, (2017). Ongoing CRISM investigation of ridge networks and their phyllosilicate-bearing host unit in the Nili Fossae and Northeast Syrtis regions, LPSC 48, Abstract #2807.
- Pascuzzo, A.C.** and J.F. Mustard, (2016). Determining the composition of various martian central mounds, LPSC 47, Abstract #2758.
- Pascuzzo, A.C.** and J. F. Mustard, R. M. Newton, (2016). Geologic mapping and characterization of Nicholson crater, Mars, LPSC 47, Abstract #2435.
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The real start down my planetary science career path was the summer internship at LPI (ca. 2014). To my LPI fam: Ellie Armstrong, Kaitlyn McCain, Stefan Farsang, Ben Go, Jeffrey Murl, Elise Harrington, Zack Torrano, Steven Dobb, Tim Gregory, and my roommate at the time—Molly Johnson—I'll never forget the office antics at the JSC and LPI headquarters. Being there with you all and growing together really inspired me to pursue graduate school in planetary science. To Zack and Steven, being able to meet up annually at LPSC these past few years (how many years has it been now!?) has been such a highlight. It's a reunion I always look forward to. Tim and Molly, I will always cherish our world record breaking 48-hour geology road trip from Houston to the Grand Canyon, Phoenix, and back to Houston. Those were truly unforgettable hours/days (P.S. I still have that clip board with our trip agenda taped to it). Last, but not least, a huge thanks to my internship mentor Carl Allen. Thanks to you I took a deep dive into Mars research. Seven years later and I'm still loving this rusty, cold, barren planet ☺.

This dissertation wouldn't have been possible without my amazing collaborators: Ali Bramson, Patricio Becerra, Shuai Li, Ray Arvidson, and Thomas Condus. Ali, you've been not only a great research collaborator but also a great friend. Working this past year on our project together has been the most fun I've had doing research. Pat, thank you for the sanity checks and data validation! Shuai, thank you for all your advice when it came to Hapke theory and in lending your RTM code. A big thanks to Ray and Thomas for hosting my visit to Wash U. for my training in using the DISORT-SSA routine. Science is not a solo endeavor. You all truly made this dissertation possible.

To the Mars Polar Science community—ya'll are a great group of people and you welcomed me with open arms. I'll never forget attending and presenting at the International Polar Science Conference in Ushuaia, Argentina. It was then that I felt truly welcome to the research community and established many long-term connections. Every Mars Scientist is a polar scientist ☺. To my Ph.D. and Prelims committee (Ralph Milliken, Jim Head, Chris Huber, Ingrid Daubar, and Wendy Calvin) and Advisor, Jack Mustard, thank you for pushing me to become a better scientist. Prelims (x2) was a struggle, but I wouldn't be the scientist I am now without that push. And a big thanks to Olga Huber for being my department mom ☺.

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Table of Contents

Title Page	
Copyright Page	
Signature Page	iii
Curriculum Vitae	iv
Acknowledgements	x
Table of Contents	xii
List of Tables	xv
List of Illustrations	xvi
 <u>Chapter 1: Climate record recorded in the martian ice caps? An Introduction.....</u>	<u>1</u>
1. Setting and motivation	2
2. Geologic context: North Polar Layered Deposits	5
2.1. Marker beds	5
2.2. Dust: sublimation lags vs veneer.....	8
3. Dissertation structure	10
References.....	12
 <u>Chapter 2: Development and evolution of outcrops of icy layers on Mars' north polar ice sheet.....</u>	<u>17</u>
Abstract	18
1. Introduction.....	19
2. Observational evidence for ice loss.....	22
2.1. Methods	22
2.1.1. Data and location details	22
2.1.2. Layer identification from stratigraphic profiles and protrusion measurements.....	25
2.1.3. Statistical correlation.....	28
2.2. Observational evidence: Results.....	29
2.2.1. Brightness and slope versus protrusion.....	29
3. Modeling ice instability.....	34
3.1. Sublimation model.....	34
3.1.1. Model set-up.....	34
3.1.2. Thermal conduction model.....	35
3.1.3. Ice loss calculations.....	36
3.1.4. Strata properties.....	38
3.2. Model Results.....	41
3.2.1. Effect of albedo and dust content of strata on sublimation and lag production rates.....	42
3.2.2. Effect of wind speed and trough topography on sublimation and lag production rates.....	43
4. Framework for forming protruding marker beds.....	44
4.1. Number of years to produce observed MB protrusion?	46
5. Discussion.....	48
5.1. Trough evolution and development of protruding marker beds	48
5.2. Spatially variable environmental conditions controlling protrusion laterally	51
6. Conclusion	51
Acknowledgments	52
Supplementary Material.....	53
S1. Modeling diffusive loss through a lag.....	53
S2. Sublimation rates used in calculating the number of year to form MB protrusion.....	54
References.....	56

Chapter 3: Sensitivity analysis of ice/dust aerosol and phase function assumptions on Hapke spectral unmixing and band depth parameters of martian water ice.....	61
Abstract	62
1. Introduction	62
2. Data set and processing.....	66
2.1. CRISM data and study site	66
2.2. Radiative transfer modeling of aerosols and surface using DISORT.....	67
2.2.1. Parameterizing the atmospheric model.....	70
2.2.2. Parameterizing the surface boundary.....	71
2.2.3. Assessing the surface single-scattering albedo (SSA) retrievals	74
2.2.4. DISORT-EPF SSA retrieval method versus simplified SSA retrieval method	75
3. Spectral analysis techniques	75
3.1. Spectral Unmixing of SSA using Hapke theory	75
3.1.1. Hapke modeling caveats to keep in mind	76
3.2. H ₂ O index band depth parameter.....	77
4. Results.....	78
4.1. Determining the best retrieval of SSA.....	78
4.2. Impact of aerosols and scattering behavior assumptions on spectral features.....	81
4.2.1. Spectral effects due to aerosol opacity assumptions.....	81
4.2.2. Spectral effects due to phase function assumptions	82
4.3. Sensitivity of Hapke Results	84
4.3.1. Sensitivity of nonlinear spectral modeling to optical depth uncertainty.....	84
4.3.2. Sensitivity of nonlinear spectral unmixing to phase function parameter uncertainty	85
4.4. Sensitivity of the H ₂ O index	86
4.4.1. Sensitivity of the H ₂ O index to optical depth uncertainty	87
4.4.2. Sensitivity of the H ₂ O index to phase function parameter uncertainty.....	88
4.5. Differences in spectral results from using EPFs (DISORT-EPF approach) versus the central scene (simplified approach).....	88
5. Interpretation of sensitivity tests and suggestions for future CRISM studies.....	90
5.1. Hapke	90
5.2. H ₂ O index	92
6. Conclusions.....	92
Acknowledgements.....	94
Appendix A. Surface scattering model equations and non-parameterized constraints used in DISORT.....	95
Appendix B. Equations used for inverse Hapke modeling of the retrieved SSA data using the optical constants of mixture endmembers	97
Supplementary Materials	99
S1. Determining the best SSA retrieval.....	99
References	104

<u>Chapter 4: Present-day surface ice/lithics properties of Mars' North Polar Layered Deposits controlled by recent post-depositional processes and regional depositional variability.....</u>	113
Abstract.....	114
1. Introduction.....	115
2. Methods.....	117
2.1. Data and study locations	117
2.2. Atmospheric and photometric correction.....	120
2.3. Nonlinear spectral unmixing	120
2.4. Statistical analysis	122
3. Results	123
3.1. Intertrough variability	124
3.2. Intratrough variability: Dust veneer.....	126
4. Discussion.....	127
4.1. Are the MBs expressing inherent layer properties or properties related to post-depositional aeolian processes within the trough?	127
4.2. Distinct veneer expression.....	129
4.3. Comparison between radar- and VNIR-derived abundance estimates.....	130
5. Conclusion.....	131
Acknowledgements	134
Supporting Materials.....	134
References.....	138
<u>Chapter 5: Summary and future suggested research directions.....</u>	145
1. Summary of findings.....	146
1.1. The effect of post-depositional processes on the spatio-temporal variability of ice cap evolution.....	146
1.2. Detangling the expression of inherent layer properties from properties reflecting modern post-depositional aeolian activity.....	147
References.....	149

List of Tables

Chapter 2

Table 1.....	25
Table 2.....	31
Table 3.....	37
Table 4.....	40
Table 5.....	41
Table 6.....	48
Table 7.....	48
Table S1.....	54
Table S2.....	54
Table S3.....	55

Chapter 3

Table 1.....	72
Table 2.....	73
Table A1.....	96
Table A2.....	98
Table S1.....	100

Chapter 4

Table 1.....	118
Table S1.....	135
Table S2.....	135
Table S3.....	136
Table S4.....	136
Table S5.....	136
Table S6.....	137
Table S7.....	137

List of Illustrations

Chapter 1

Figure 1.....	4
Figure 2.....	7
Figure 3.....	10

Chapter 2

Figure 1.....	24
Figure 2.....	27
Figure 3.....	28
Figure 4.....	32
Figure 5.....	33
Figure 6.....	40
Figure 7.....	43
Figure 8.....	44
Figure S1.....	53
Figure S2.....	53

Chapter 3

Figure 1.....	67
Figure 2.....	68
Figure 3.....	73
Figure 4.....	74
Figure 5.....	80
Figure 6.....	83
Figure 7.....	85
Figure 8.....	87
Figure 9.....	89
Figure 10.....	91
Figure S1.....	100
Figure S2.....	101
Figure S3.....	102
Figure S4.....	103

Chapter 4

Figure 1.....	119
Figure 2.....	122
Figure 3.....	124
Figure 4.....	125
Figure 5.....	126
Figure S1.....	134

Chapter 1

Climate recorded in the martian ice caps | An introduction

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1 Setting and motivation

The geologic history of Mars is divided into three epochs: the Noachian (>3.7 Ga), the Hesperian (~ 2.9 – 3.7 Ga), and the Amazonian (~ 2.9 Ga–present). The Amazonian epoch describes most of Mars' geologic history and has been relatively quiescent concerning large-scale geological (e.g., volcanic) and climatic (e.g., fluvio-lacustrine) activity. In contrast to the wetter Noachian and Hesperian epochs, the Amazonian is characterized by cold and dry conditions—dominated by aeolian and ice processes on a global scale (e.g., Diniega et al., 2021). The Amazonian climate has been mainly controlled by quasi-periodic variations of Mars' orbital and rotational axis parameters—obliquity, the argument of perihelion, and eccentricity—similar to Earth's Milankovitch cycles (e.g., Toon et al., 1980). These variations affect the average solar flux reaching Mars' surface and drive the instability of ices, thus the global spatial and temporal distribution of ice reservoirs. In addition to these astronomical forcings, the Amazonian climate is also influenced by the martian dust cycle, which is poorly understood (Newman et al., 2021).

Analogous to ice sheets and ice cores on Earth, the icy and dusty beds of the North and South Polar Layered Deposits (PLD) on Mars likely record information on the climatic and atmospheric states during layer accumulation, thus potentially recording global climate variations due to astronomical forcings. In this thesis, I focus on the North Polar Layered Deposits (NPLD) exposed within these troughs. Since its discovery in the 1970s, the ~ 1500 -m-thick dome, comprised of distinct bright- and dark-toned layered ice, is hypothesized to record variations in relative abundances of water ice and atmospherically-deposited dust over the past ~ 4 Myr due to astronomically-forced cyclical changes in solar insolation (Murray et al., 1972; Cutts and Lewis, 1982; Howard et al., 1982; Laskar et al., 2002; Tanaka K. L. et al., 2003; Milkovich and Head, 2005; Tanaka, 2005; Levrard et al., 2007; Perron and Huybers, 2009; Putzig et al., 2009; Hvidberg et al., 2012; Becerra et al., 2017). Although there are not yet ice cores from the martian PLDs, various remotely-sensed data have been collected of the deposits, including shallow orbital sounding radar observations and visible and near-infrared observations of outcrops exposing ~ 500 m of ice and dust-rich strata within troughs that dissect the ice caps (Figure 1). Various research has attempted to constrain the amount of time represented and recorded by the exposed NPLD. Laskar et al. (2002) predicted that the upper 300 m of the NPLD likely accumulated over the past 400 kyr. Work by Smith et al. (2016) supports this prediction,

suggesting that the upper 320 m likely accumulated over the past 370 kyr. Other researchers have suggested that the upper ~500 m between 250 kyr–1 myr (Hvidberg et al., 2012).

Despite this progress over the past few decades, our understanding of how NPLD strata are specifically linked to climatic and atmospheric states is muddled. As a result, many outstanding questions remain, including 1) what timespan do specific layers or packets of layers record?; 2) how complete is the record (i.e., what events and processes are recorded in the layers, and how are these records erased)?; and 3), what is the temporal resolution of the record in these deposits? Today's challenge in addressing these questions is partly because the ongoing gross and net annual exchange of H₂O onto/out of either cap is unknown. The present-day mass balance of volatiles is required for tuning and validating global and regional mesoscale present and past climate models, which aid remote sensing investigations in linking climate to the stratigraphic record.

Interpreting the climate record recorded in the stratigraphy is a two-step process that involves: 1) an accurate description of the stratigraphy by integrating and analyzing multiple datasets (i.e., radar, visible, hyperspectral); and 2) use of the stratigraphic record to constrain climate and layer accumulation model inputs for stratigraphic reconstruction (e.g., Levrard et al., 2007; Hvidberg et al., 2012; Hvidberg, 2016). Unfortunately, current accumulation models fail to model the observed stratigraphy. This implies that our current description of strata interpreted to reflect global climate variation is incomplete. Furthermore, our current description of the NPLD strata is complicated by the geographically-variable aeolian surface processes and atmospheric conditions across the polar cap (e.g., Howard, 2000; Smith and Spiga, 2018).

Spatial variability in solar ablation, wind transport, and atmospheric deposition of ice and dust governs the change in the mass balance of polar ice deposits. The variability of these processes affects both primary deposition and preservation layers on local scales of 10's to 100's of km—as indicated by both subsurface radar (Smith and Holt, 2010; Smith et al., 2013; Smith and Holt, 2015; Bramson et al., 2019; Lalich et al., 2019) and surface observations (e.g., Tanaka et al., 2008; Becerra et al., 2016) of the NPLD strata. Paleo post-depositional modification of recently deposited layers affects whether a layer ends up being preserved. Modern post-depositional surface modification (such as erosion and deposition of dust veneers) can affect our climatic interpretation of beds that are now exposed within troughs by modifying their surface

composition and texture (Herkenhoff et al., 2007; Rodriguez et al., 2007; Fishbaugh et al., 2010a). Becerra et al. (2016) correlated regionally continuous beds (implying cap-wide accumulation of such strata) between multiple spiral trough exposures. However, they noted multiple locations of stratigraphic unconformities, suggesting that although some beds are continuous between one or two troughs does not mean they are (or ever were) cap-wide. Additionally, the visually-resolvable layers between regionally continuous beds vary from trough-to-trough and are uncorrelatable—consistent with previous identifications of NPLD strata that pinch out (Tanaka et al., 2008). Becerra et al. (2016) suggest that local conditions during ice and dust deposition result in differences in what is ultimately recorded. This is supported by the observed variability in the slope of proximal trough migration paths, where radar discontinuities in stratigraphy highlight trough migration via sublimation vs. accumulation as a function of time (Smith and Holt, 2010; Smith et al., 2013; Smith and Holt, 2015; Bramson et al., 2019). Additional evidence from radar observations shows a cap-wide radar reflective layer with dust abundance estimates ranging between 15–40%, supporting the idea that large-scale, regional and small-scale, local processes control the geographic distribution of ice and dust deposition and preservation (Lalich et al., 2019).

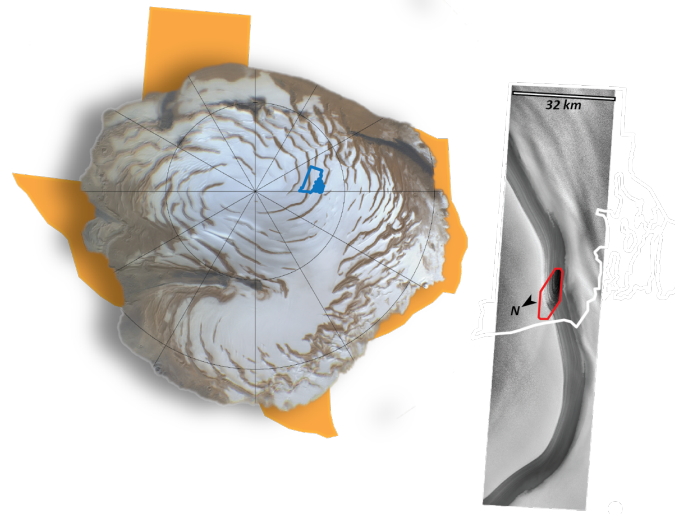


Figure 1. Left: Planum Boreum, compared to the size of Texas (orange) and Rhode Island (blue). The spiral troughs are naturally highlighted by the darker tone of the NPLD strata exposed by the dissecting troughs. Image clipped from Mars Color Imager (MARCI) raster from Tanaka and Fortezzo (2012). Right: Subset image zoomed into the spiral trough covered by the Rhode Island outline in the left image. Data shown on the right is a full, greyscale Context Camera image (CTX) acquired during northern summer. CTX observations take ~32 km wide swaths. The red polygon outlines the first orthorectified image produced over NPLD stratigraphy exposed within the dissecting trough. This location is known as site N0 and its one of two sites investigated in this dissertation.

This brings us to the motivating theme of the work presented in this dissertation. A better characterization of the processes and timescales driving NPLD surface modification is required in order to interpret the environmental records preserved within troughs—including how the primary deposition of the NPLD is linked to climate. Characterizing present-day polar cap surface modification processes on local scales (opposed to cap-wide) is an indispensable step in deciphering the cap's complex evolution and environmental record. Surface modification processes of interest in this dissertation are aeolian erosion of troughs, sublimation (and lag formation) of ice, redistribution of dust and ice, and allochthonous dust-veneer deposition and induration within spiral troughs. By investigating the spatial variability in the NPLD surface expression in spiral troughs, I begin to untangle the record of modern surface processes and volatile exchange from the paleo record.

2 Geologic context: North Polar Layered Deposits

2.1 Marker beds

Earlier studies revealed a variety of surface textures and erosional styles throughout the vertical NPLD sequences: 1) dark-toned, regionally-traceable, protruding, knobby, and pitted marker beds (MBs – Figure 2b & 2c), 2) brighter-toned undifferentiated and topographically recessed beds, and 3) thinly-layered sets with variable topographic expression (Malin and Edgett, 2001; Milkovich and Head, 2006; Fishbaugh and Hvidberg, 2006; Herkenhoff et al., 2007; Fishbaugh et al., 2010a, 2010b). These studies suggested that the erosional morphology, apparent brightness, and protruding nature of a layer (or set of layers) is a function of the ice-to-dust ratio driven by the insolation cycle.

Other studies of the NPLD (Herkenhoff et al., 2007; Fishbaugh et al., 2010a) suggest that the morphology and brightness of the layering could be a function of seasonal wind and climate patterns controlled by trough topography. Fishbaugh et al. (2010) were the first to qualitatively show that the MBs appear rougher in the HiRISE Digital Terrain Model (DTM) hillshade map while the interbeds (i.e., strata that are not marker beds) appeared smoother. These MB strata protrude from the trough wall relative to the lighter-toned interbeds that typically sandwich them (Fishbaugh et al., 2010a; Becerra et al., 2016). Their

ablation-resistant nature and roughness are thought to be due to higher dust content and protective, indurated, or loose surface lag.

The formation of marker beds in relation to climate and surface processes recorded at the time of deposition remains debated. There are two competing hypotheses for MB formation:

1. High dust and ice accumulation simultaneously (low solar insolation at the north pole) resulted in the dark-toned layers representing initial depositional composition (Putzig et al., 2009; Laskar et al., 2002).
2. MBs are themselves a build-up of sublimation lag formed when the climate favored deposition of dirty ice and subsequent sublimation annually (Levrard et al., 2007).

Obtaining the inherent lithic-to-ice ratio of MBs from multiple data sets (radar and hyperspectral) would help us infer which formation mechanism is more likely. Additionally, studying MBs' present-day topographic expression can inform us of recent sublimation rates and the surface/atmospheric conditions that drive spatial variability of ice loss and the spatial scale and timescales of ice loss within troughs.

Fishbaugh et al. (2010) were also the first to create a stratigraphic column with designated nomenclature (MB $\pm n$ where MB0 is the first marker bed identified by Malin and Edgett 2001). Motivated by their work, Becerra et al. (2016) investigated NPLD stratigraphy at 16 different trough sites (black and blue dots in Figure 2a) across the polar cap determined to find a sequence of beds that could be used to link and infer global climate. Through protrusion-based correlation analysis, they successfully correlated six of sixteen stratigraphic profiles (trough sites N0, N1, N6, N8, N10, N15; sites N0, N8, and N15 shown in Figure 2). MBs beds identified at these different locations are laterally continuous and have been correlated across these two troughs and between trough sites (Becerra et al., 2016). This correlation implies that "main sequence layers" (layers that most likely record global climate change (Becerra et al., 2017)) are exposed at these locations. They define the main sequence to be composed of layers MB8 through MB-5; however, not all the correlated sets contained this entire set either due to missing layers from suggested local unconformities or simply due to the depth of the spiral trough.

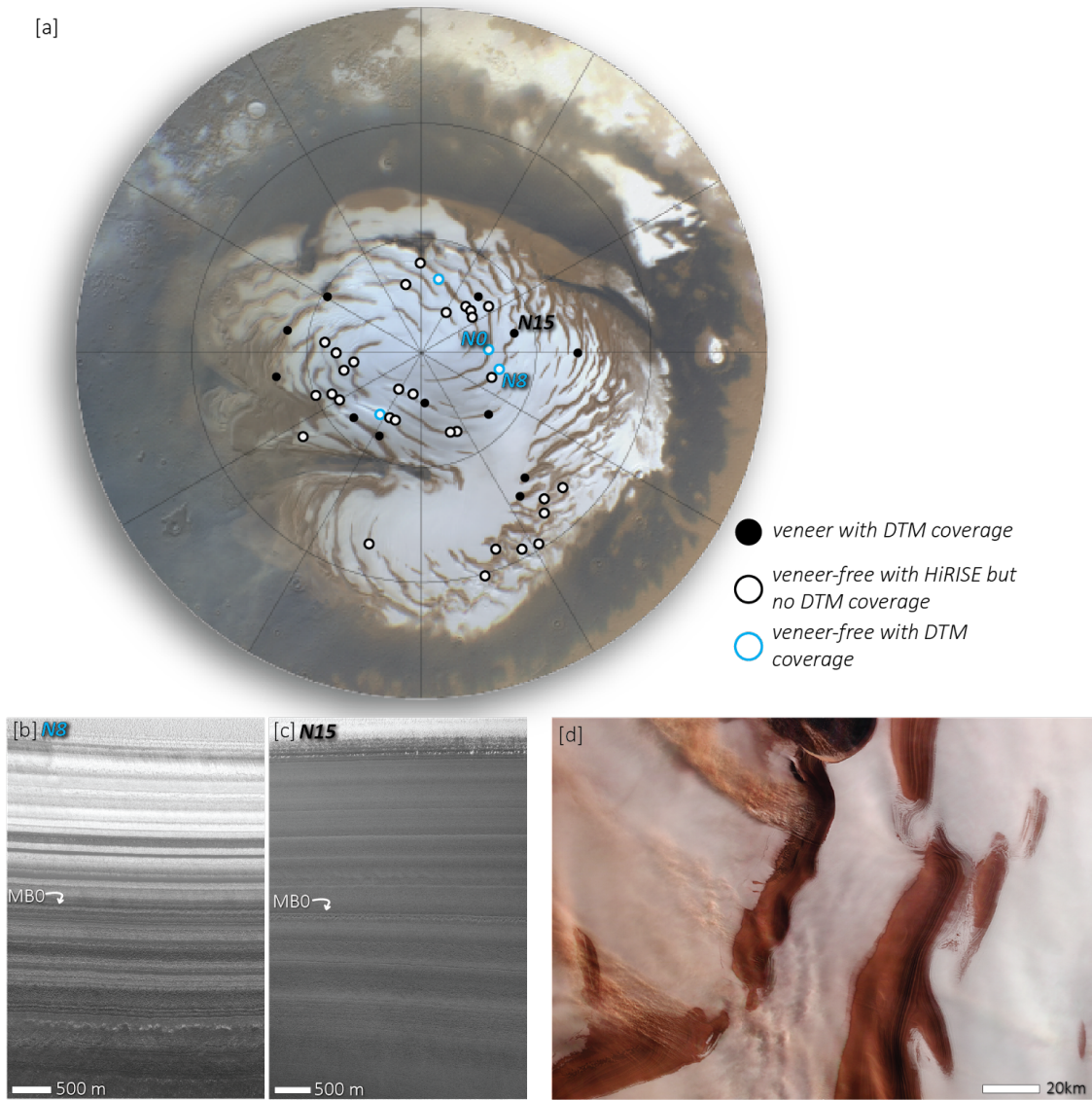


Figure 2. [a] MARCI true-color map of the north polar region (Tanaka and Fortezzo, 2012). The spiral troughs are naturally highlighted by the darker tone of the NPLD strata exposed by the dissecting troughs. Dots indicate equatorial facing trough walls that have HiRISE coverage acquired in northern martian summer between Earth years 2006-2017. Solid black dots indicate locations where there is HiRISE DTM coverage over a veneered trough wall section. Black dots also happen to be 12 of the 16 locations Becerra et al. (2016) investigated. The majority of the equatorial-facing trough wall are veneered. Only small sections of troughs are veneer-free. Black outlined white dots are veneer-free trough wall NPLD outcrops with HiRISE coverage but no DTM coverage. Blue outlined white dots are veneer-free NPLD outcrops with HiRISE DTM coverage. The four blue outlined dots are the remaining 4 of 16 sites Becerra et al. (2016) investigated. [b] HiRISE image of veneer-free spiral trough wall at site N8 (HiRISE image: ESP_019210_2665). [c] HiRISE image of veneer covered spiral trough wall at site N15 (HiRISE image: PSP_009980_2660). [d] HRSC color composite image showing strong winds perpendicular to trough length during northern summer. These strong mid-summer winds appear to transport dust from the veneered slopes non-uniformly. Image Credit ESA/DLR/FU Berlin, CC BY-SA 3.0 IGO. Orbit #3670.

2.2 Dust: sublimation lags vs. veneers

Herein, I define the difference between sublimation lags versus veneers when discussing polar ice deposits. Over the past few decades, a variety of names have been used to describe the (non-dune) sedimentary material distributed across the caps, including dark mantling deposits, dark mantles, young mantling deposits, dark sedimentary deposits, north polar veneers, dust veneers, dust mantle, regolith, dark interior deposits, dust lag, depositional lags, lithic debris, and residual dust.

Sublimation of 'dirty' ice leaves behind a thin layer of the lithics once embedded within. The amount of lag accumulated on the surface is proportional to the sublimation rate and volume percentage of lithics. On Mars, the majority of the lithics contained in ice are believed to be compositionally that of martian airfall mineral dust (Pollack et al., 1979; Howard et al., 1982). Martian mineral dust is similar in composition to martian regolith, which is similar in composition to altered basalt (e.g., palagonite) and globally homogeneous (Bell et al., 2000; Berger et al., 2016). Herein, I use the terms lithics and dust interchangeably. It is unclear whether the surface of the exposed MB is a sublimation lag or if it is the freshly exposed surface of a MB as they can be very thin (μm – cm) and likely visually have similar texture to the underlying MB. With this in mind, it is possible that marker beds may be "protected" by a sublimation lag and that the surface composition of MBs may or may not reflect inherent internal ice-to-lithic ratio of marker beds, but instead that of a sublimation lag. Here, when using the phrase sublimation lag, I strictly relate it to the implied autochthonous nature of lag formation. Therefore, if a lag is subsequently eroded and redeposited elsewhere, it is no longer called a sublimation lag.

North polar veneers are discontinuous, dark, sedimentary mantles (Figure 2c) found along the equatorial-facing slopes and at the bottom of polar troughs. They can be both immobile and highly mobile over annual timescales depending on the relative age of veneer emplacement—older ones are proposed to be thicker and/or indurated by pore-filling ice or salts (Rodriguez et al., 2007). The spatio-temporal distribution and thickness of veneers over seasonal and annual timescales would provide important information to their present-day impact on the surface exchange, or lack thereof, of H_2O . Such information would allow the first steps needed to more confidently extrapolate the behavior of paleo-veneers' (which are not preserved in the polar record as we know it) contribution to recent trough migration/excavation.

Presently, veneers are observed partially overlying the residual ice cap implying their emplacement is younger than the topmost residual ice layer of the cap. Figure 2a maps the locations of HiRISE data (imaged during northern summer between Earth years 2006–2017) with coverage of equatorial-facing trough walls and whether an imaged portion of the trough wall is veneer-free or blanketed by a veneer. Figure 2b–2c showcases two different HiRISE observations within the same spiral trough (~90 km from one another) with opposing veneer coverage. Veneers have tapered and diffuse margins similar to windstreaks, which suggests they are allochthonous aeolian deposits and not in-situ trough wall sublimation lags. Rodriguez et al. (2007) propose that the source material of veneers is another discontinuous, but unconsolidated, sedimentary deposit composed of sand-sized lithics, which is highly mobilized by wind (possibly migrating sand sheets), found at the base of the pole-facing slopes of many troughs and in some case climb the pole-facing trough slopes (Figure 3c). These possible veneer source deposits are called "intermediate deposits" (Rodriguez et al., 2007; Tanaka et al., 2008). Predicted wind regimes across inner Planum Boreum—from mesoscale atmospheric circulation modeling (Smith and Spiga, 2018)—suggests the veneer in a trough (Figure 3d) would be sourced from intermediate deposits residing in troughs that located at higher latitudes (Figure 3c). The origin of the intermediate deposits has not been extensively studied. However, possible sediment sources include NPLD lithics eroded from the equatorial-facing slopes of troughs (Figure 3b) and/or atmospherically deposited (and aggregated) dust which settles to the surface and is transported to topographic lows and towards the peripheries of the ice cap after the winter polar vortex dissipates in northern hemispheric spring (Figure 3a) (e.g., Calvin et al., 2015). The albedo of veneers varies, possibly due to thickness and age, and contain water ice absorptions of varying strength upon inspection of VNIR spectra (Rodriguez et al., 2007).

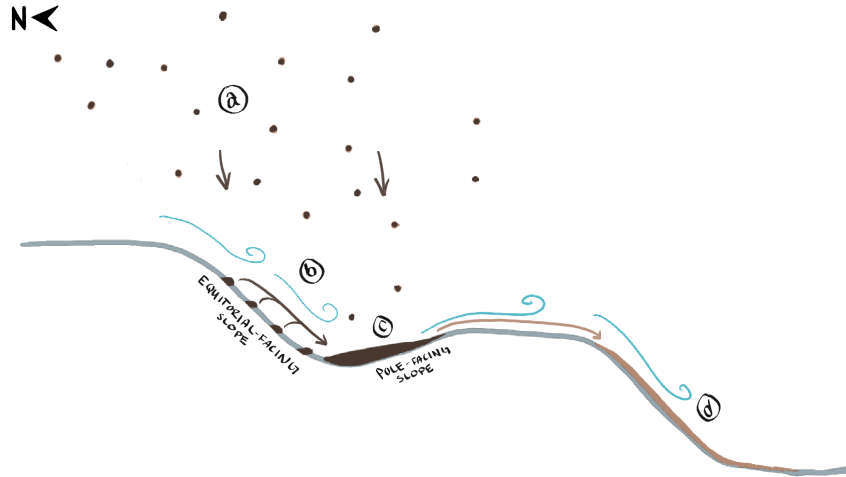


Figure 3. Cartoon illustration showing the hypothesized source material (i.e., (c) intermediate deposits) for "dust" veneers (d) as well as the two possible sources for the intermediate deposit lithics—(a) dust settled from the atmosphere from dissipation of polar vortex and/or (b) NPLD lithics eroded from the equatorial-facing walls of troughs.

3 Dissertation structure

The two main questions addressed in this dissertation are:

- 1) What are the timescales, spatial scales, and magnitudes of water vapor leaving the Mars' north polar spiral troughs, and how have post-depositional surface conditions affected these over the past 300 kyr?
- 2) What are the surface properties of main sequence marker beds of the NPLD and the young, discontinuous dust veneer deposits that mantle the vast majority of equatorial-facing trough walls present-day and do marker bed properties reflect inherent layer properties associated with primary accumulation conditions of the layer or do they reflect post-depositional aeolian activity?

In Chapter 2, titled "Development and evolution of outcrops of icy layers on Mars' north polar ice sheet," I address Question 1 to determine the effect of post-depositional processes on cap evolution. I do this using high-resolution topography and novel geomorphometric techniques to characterize spatial variability in trough wall retreat. Then, I use thermal conduction modeling to investigate ice stability of different NPLD strata types (marker beds and interbeds) to 1) determine whether differential sublimation can produce the characteristic topographic signatures of marker beds and 2) what (combination of) environmental factors drive heterogeneous NPLD retreat within spiral troughs?

Chapter 3 and 4 combined addresses Question 2. Chapter 3 is precursor work which leads into spectral analysis performed in Chapter 4. Chapter 3 addresses the sensitivity of propagated uncertainty due to assumptions made in data processing and conversion from reflectance to single-scattering albedo (SSA) into the spectral analysis of Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) data of icy-dusty polar terrains. The following questions are addressed in Chapter 3. First, how do untreated aerosols affect the spectral characteristics and interpretation of the ice-rich surface through typical spectral techniques such as H₂O Index parameters and Hapke mixture modeling? Second, how does assuming isotropic versus anisotropic scattering affect spectral characteristics and interpretation of the ice-rich surface (via H₂O Index and Hapke mixture modeling)? How does using a Discrete Ordinate Radiative Transfer Modeling (DISORT)—which utilizes the full viewing geometry of CRISM emission phase function (EPF) observation—to retrieve SSA affect the spectral results compared to the more common and simplified SSA retrieval methods that only use I/F (radiance at the sensor (I) divided by F, which is the incident solar irradiance divided by π) and viewing geometries from the near-nadir, central CRISM image?

In Chapter 4, I take a hyperspectral approach to detangling the expression of inherent layer properties from properties reflecting post-depositional aeolian activity. I use Hapke nonlinear spectral unmixing of DISORT-EPF processed CRISM SSA data at two different NPLD spiral trough outcrops to determine whether the surface properties of MBs are primarily a function of past environmental conditions during layer accumulation or recent post-depositional aeolian processes. Half of one of the NPLD outcrops is blanketed by the north polar dust veneer which we determine the properties of. This work will help future studies detect climate signals more confidently, by accurately describing and teasing inherent properties of the layers that make up the climate record from modern, erosional surface properties as inherent layer properties most closely relate to the environmental conditions at the time of layer accumulation.

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Chapter 2

Development and evolution of outcrops of icy layers on Mars' north polar ice sheet

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Abstract

We investigate the processes and factors within spiral troughs that contribute to the evolution of the north polar cap on a local km-scale to gain insight into the recent amounts of sublimation and timescales of active trough wall retreat. This was done through a combined approach of geomorphic characterization of trough wall topography signatures (using HiRISE DTMs and novel approach to calculating true layer protrusion) and thermal conduction modeling. We found that the topographic relief of protruding marker beds (MBs) decreases as a function of decreasing trough slope and brightness (used as a proxy for dust cover). We model the contribution of insolation-induced sublimation on two different stratum types (higher dust content marker beds and lower dust content interbeds) under a range of albedo and wind speed conditions and used the model outputs to develop a framework for developing marker bed protrusion determined. Our modeling and theoretical calculations imply that MBs can develop into 1–5 m protrusions on the order of 1000's of years via cyclical bursts of differential sublimation modulated by lag removal and variable lag production (a function of dust content and sublimation rate). We determined that the difference in dust content between a MB and the beds that sandwich it is important in causing the MBs to form as protruding layers. Higher dust content results in quicker formation of lag on the MBs. However, the lag does not need to be resistant to mechanical erosion to cause a MB to protrude. Under a range of possible surface albedos, dust content of strata, and wind speed down the equatorial-facing trough walls, we find that MB topographic relief can easily be formed within a single martian precession period (e.g., high north polar insolation lasting 25,000 years driven by Mars' axial precession). If the present-day protruding and recessive signature of exposed NPLD strata was strictly driven by differential sublimation (no mechanical erosion of ice or erosionally resistance lags) we find that the dominant amount of ice retreat would have occurred as early as 60–125 kya. However, due to the speed at which topographic relief can form (1–20 kyr) leads us to suggest that thick insulative allochthonous dust veneers would be needed to force hiatuses in trough wall retreat during high polar insolation periods. Lateral variability in ice loss, and thus present-day protrusion variability, is best explained by the variability in slope wind speed and surface dust removal rate and their combined relationship with the differential sublimation and lag production rate between MB and interbeds (IB).

1 Introduction

One of the main open issues in Mars cryospheric science is the characterization of the mass balance of the polar ice deposits, a critical step to understanding the current surface-climate interactions, and thereby interpreting the climate record preserved in the ice (Becerra et al., 2021, *in press*). Current knowledge of whether Mars' North Polar Layered Deposits (NPLD) experience growth or loss due to H₂O and dust exchange between the surface and atmosphere remains insufficient. Transport of water and dust onto/out of either polar cap is variable spatially and temporally on seasonal to decadal and longer timescales, and the processes driving this variability are not well understood. Characterizing present-day polar cap surface modification processes is an indispensable step in deciphering the cap's complex evolution, and reveal the environmental records preserved within.

The spiral troughs of Planum Boreum (Figure 1), which dissect the NPLD, play an important role in shaping their evolution and the record of past martian climate stored within. These spiral troughs expose layers of water ice and mineral dust mixtures with properties that vary geographically and with depth (e.g., in their surface expression, ice-to-dust ratio, and regional continuity). Over the past few decades, there has been an ongoing debate about how these spiral troughs form. One side of this debate supports the formation via erosion after NPLD deposition (Weijermars, 1986; Ivanov and Muhleman, 2000; Kolb and Tanaka, 2001; Ng and Zuber, 2006; Zeng et al., 2008; Rodriguez et al., 2021). On the other side is the more widely accepted mechanism of constructional migration—which involves simultaneous deposition and ablation (Squyres, 1979; Howard et al., 1982; Fisher, 1993; Howard, 2000; Smith et al., 2013). Studies have shown troughs experience different initiation points in space and time, as well as horizontal migration through time as obliquity cycles force the north polar region to fluctuate from periods during which ice is and is not stable (Smith and Holt, 2010; Smith et al., 2013; Smith and Holt, 2015; Bramson et al., 2019). This has led to further interest in understanding the mechanisms and processes behind trough evolution, including the effects of katabatic winds and time-variable thickness of surficial dust (Smith and Spiga, 2018; Bramson et al., 2019). Thus, further evidence on the exchange of volatiles as a function of climate and how fluctuations in this exchange are expressed within and on the NPLD may help resolve this issue.

Variability in solar ablation, wind transport, and atmospheric deposition govern the change in mass balance of polar ice deposits, and thereby on the spatial and temporal changes in trough migration. More

specifically, regional and local conditions can cause ablation and sublimation rates to vary spatially within the local environment of a trough (Smith and Holt, 2015). Spatial variation in sublimation can result from the combined variability in trough topography, albedo (e.g., due to dust cover or ice grain size), and thermal inertia (e.g., due to changes in surface material, grain size, porosity). Firstly, exposed NPLD layers are observed to pinch out occasionally. These unconformities have been mapped (Tanaka et al., 2008; Becerra et al., 2016), thus suggesting that some locations and periods in time experienced either differences in depositional and/or erosive and abrasive processes. Radar observations from the Shallow Radar (SHARAD; (Seu et al., 2004)) that show past trough locations as unconformities in the subsurface stratigraphy (known as trough migration paths), suggest both temporal and spatial variability in accumulation and sublimation rates (Smith and Holt, 2010; Smith et al., 2013; Smith and Holt, 2015; Bramson et al., 2019). Additional evidence from radar observations shows a cap-wide radar reflective layer with dust abundance estimates ranging between 15–40%, supporting the idea that both regional and large-scale processes control the geographic distribution of ice and dust deposition (Lalich et al., 2019). Recent modeling of trough migration via sublimation/ablation of the equatorward-facing trough slope suggests that an appreciable long-term dust lag or veneer must be present on the floor of the troughs for the modeled migration paths to match those observed in the SHARAD data (Bramson et al., 2019). A thick (cm-scale or thicker) dust veneer or lags inhibits sublimation through thermal insulation, while thermally thin veneers or lags (<mms) can promote sublimation by decreasing the albedo of the surface (Bramson et al., 2019). Despite these insights, the conditions and processes that occurred on Mars which led to the current polar ice cap structure and morphology are not well understood. Outstanding questions include: how can the various surface conditions that affect the surface expression and volatile cycle be quantified? How can the topographic expression of layers exposed at the surface today inform us about long-term dust and wind conditions?

To interpret the environmental records preserved within troughs—including the primary deposition of the NPLD—the processes and timescales behind surface modification and trough migration require better characterization. Understanding the spatial scales of the changes in water ice sublimation in troughs is critical for deciphering the polar cap’s complex evolution. Furthermore, understanding the processes that drive this heterogeneity and the modification of the present-day surface is essential to interpret the environmental records preserved within the NPLD. The goal of this study is to investigate the processes and factors within

spiral troughs that contribute to the evolution of the north polar cap on a local scale in order to gain insight into the recent (present-day to 100s kyr) rates of sublimation and timescales of active trough wall sublimation.

This paper is divided into two parts. In the first part (Section 2), we explain our use of images and digital terrain models (DTMs) from the High Resolution Imaging Science Experiment (HiRISE; (McEwen et al., 2007)) to geo-morphometrically investigate dark-toned, regionally traceable layers—known as marker beds—(Malin and Edgett, 2001; Milkovich and Head, 2005; Fishbaugh and Hvidberg, 2006; Fishbaugh et al., 2010a, 2010b; Becerra et al., 2016) exposed along the equatorial facing trough walls. Marker beds (MBs) are characterized by their ablation-resistant nature which results in them protruding from the trough wall relative to the sandwiching lighter-toned interbed layers (i.e., all layers that are stratigraphically between MBs) (Fishbaugh and Hvidberg, 2006; Fishbaugh et al., 2010a, 2010b; Becerra et al., 2016). Measuring MB protrusion can be used to understand local variation in surface conditions that may have led to this variability in topographic relief. Lateral and horizontal variation in layer protrusion can occur due to several reasons such as the amount of insolation, the layer's dust content/erosional resistance, average wind speed (temperature gradients and slope), exposure/resurfacing age of the layer—if varied spatially (i.e., how long has it been subject to sublimation and aeolian surface modification), and longevity and thickness of post-depositional trough wall dust-veneers. Thus, in the first part of this paper, we determined how the topographic expression of layers varies laterally.

However, from observational data alone, we cannot disentangle the causes of the variable protrusion (e.g., whether the driving factor is the difference in insolation due to trough slope/orientation or the persistent dust presence). Additionally, it is still unknown how and why MBs protrude. Their ablation-resistant nature is thought to be due to a higher dust content and a protective, indurated dust lag; however, this has yet to be tested. In part two of this study, we implement modeling to investigate 1) whether you can achieve MB protrusion through differential sublimation between MBs and interbeds (i.e., can protrusion be achieved without accounting for erosional resistance) and 2) how trough wall slope, orientation, and dust cover—and thus, albedo—may contribute to the spatially nonuniform ice sublimation and thus layer expression variability. We synthesize (Sections 4 and 5) these results to discuss a possible framework for 1) why and

how MBs protrude in the first place; 2) how long it would take for the lateral variation in layer expression to occur as observed in part 1.

2 Observational evidence for nonuniform ice loss

2.1 Methods

2.1.1 Data and location details

This study uses Planetary Data System(PDS)-released orthorectified HiRISE stereo images (0.25 m/pixel) (McEwen et al., 2007) and HiRISE Digital Terrain Models (DTMs) (~1 m/pixel) for topography to identify MBs and their boundaries following similar layer identification methods as (Becerra et al., 2016, 2019) (layer identification described in Section 2.1.2.). Context Camera (CTX) (Malin et al., 2007) images are used as basemaps and to trace layers between HiRISE footprints. Table 1 lists the details for the HiRISE images, including date and time of acquisition, geographic coordinates, and resolution used at each location.

Secondary HiRISE products—consisting of downsampled (100 m/pixel) slope and aspect maps and high-resolution brightness maps (0.25 m/pixel)—were produced in ESRI ArcMap to assess how MB protrusion changes with respect to local slope, trough wall orientation, and local brightness (average albedo of the MB plus the corresponding interbeds (Figure 2)). Downsampling of the elevation data was performed because pressure gradients and air mass flow, and thus katabatic wind speeds, occur over length scales of a hundred or few hundreds of meters to a couple of kilometers on Earth and similarly on Mars (Rafkin et al., 2001; Petrosyan et al., 2011; Baines, 2015). The downsampled slope and aspect maps were produced by downsampling the elevation data (i.e., the DTM) from 1 m/pixel to 100 m/pixel first prior to slope and aspect calculation. To make the brightness maps, the orthorectified HiRISE images (0.25 m/pixel) were converted from digital numbers to I/F (radiance at the sensor (I) divided by the incident solar irradiance divided by π (F)), which outputs the photometrically corrected surface brightness.

The geomorphometric analysis portion of this work focuses on two spiral troughs sites (N0, N8) located ~85°N on the north polar cap (Figure 1). These trough sites were chosen for several reasons:

1. Their proximity to each other allows for higher confidence in layer traceability between troughs (Figure 1). N0 is a mere ~55 km northeast of N8 in an adjacent trough.

2. These sites are two of the six sites that Becerra et al. (2016) found to have matching layers. Specifically, the marker beds labeled by these authors as MB8, 6, 5, 3, 2, 1, 0, -1, and -2 were confirmed matches exposed in all three troughs.
3. As shown in Figure 1, each location has a different degree of dust coverage as well as differences in slope and aspect (orientation of trough relative to north) of the trough wall. Site N0 is unique in that half of its NPLD exposure is covered by a dust veneer (to the west). In addition, the slope and aspect of the trough wall change from east to west (Figure 1d–1f). Meanwhile, N8 has a fairly consistent slope and aspect laterally (Figure 1e–1g) and is completely free of a dust veneer (Figure 1c).

This variability in trough wall exposures allows us to examine how the same layer in different locations changes its topographic and morphologic expression due to external factors such as dust cover and large-scale topography. These factors affect the amount of solar radiation that reaches the ice and the local pressure and gradient-driven winds such as katabatic winds.

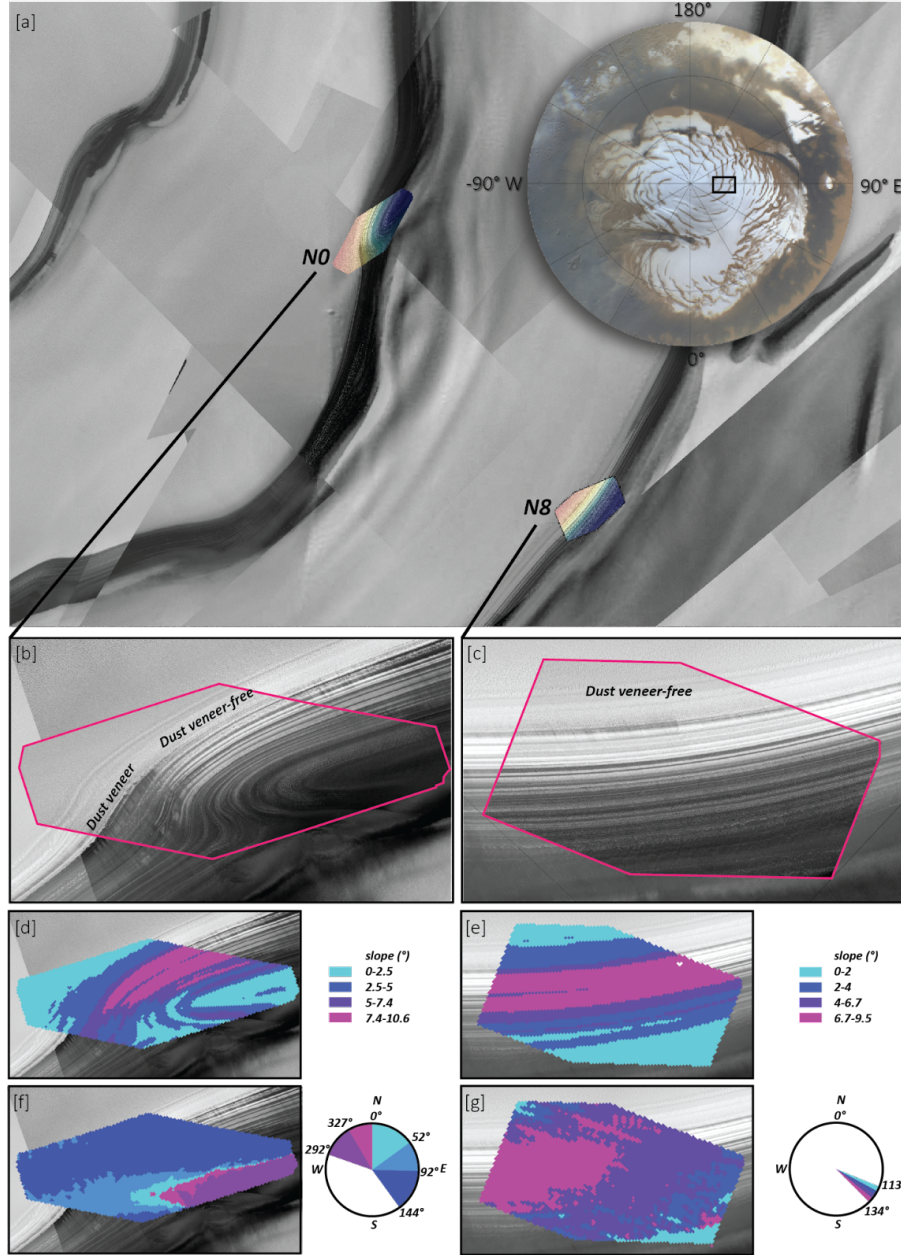


Figure 1. [a] Context image showing sites N0 and N8, the HiRISE footprint locations at each site, all shown on CTX mosaic basemap (Tanaka and Fortezzo, 2012). [b] and [c] display annotated orthorectified HiRISE images pointing out the differences in lateral dust veneer cover at the two sites. [d]–[g] show the 100 m/pixel slope maps [d and e] and aspect/trough orientation maps [f and g] made from the HiRISE DTMs at both sites. Slope and aspect binning was determined automatically in ArcMap via the data's histogram information (binning based on percent of range to overall data distribution).

Table 1. Data details per location and trough descriptions.

N0			N8	
<i>Lat, Long</i>	87.1°N, 92.7°E		86.5°N, 80°E	
<i>HiRISE DTM IDs</i>	DTEPC_001738_2670_001871_2670_U01		DTEPC_019210_2665_019158_2665_A01	
<i>HiRISE DTM resolution and vertical precision</i>	0.98 m/pixel, 1m		1 m/pixel, 1 m	
<i>HiRISE stereo pairs IDs</i>	<i>left</i>	<i>right</i>	<i>left</i>	<i>right</i>
	PSP_001738_2670	PSP_001871_2670	ESP_019210_2665	ESP_019158_2665
<i>Resolution</i>	0.25	0.25	0.25	0.25
<i>Time of data acquisition (UTC)</i>	2006-12-09 Time 21:33:26.75	2006-12-20 Time 06:12:12.94	2010-09-01 Time 09:04:57.47	2010-08-28 Time 07:49:58.44
<i>Data acquisition in Mars Year and Ls</i>	MY 28, L _s = 147.93	MY 28, L _s = 153.25	MY 30, L _s = 141.65	MY 30, L _s = 139.64

2.1.2 Layer identification from stratigraphic profiles and protrusion measurements

We use the layers identified by Fishbaugh et al. (2010b) and Becerra et al. (2016) at site N0 and by Becerra et al. (2016) at N8. We use the same layer nomenclature as these authors for consistency (e.g., MB±n). The regionally continuous layers found at both sites—or what Becerra et al. (2016) call Main Sequence layers—are MB8, MB6, MB5, MB3, MB2, MB1, MB0, MB-1, MB-2. This study focuses on these layers, specifically, due to their confirmed cross-trough correlation.

The workflow of layer identification and protrusion measurement (Figure 2, Figure 3) is as follows:

1. Produce average protrusion profiles from elevation profiles (procedure described further below).

2. Use the average protrusion profile at each site to identify which peaks correspond to which MBs, performed by cross-referencing with Becerra et al. (2016) and by visual validation through inspection of HiRISE images and hillshade products with a 10 m contour overlay in ArcMap.
3. Record the elevation for each protrusion peak and corresponding upper and lower recessive points of the strata that bound the protruding MB (protrusion minima). In addition, each layer's upper and lower boundaries were cross-referenced with those designated by Becerra et al. (2016).
4. Calculate true layer protrusion, shown as the distance between P and P' in Figure 3d (described further below). This is different from the normalized protrusion shown in protrusion profiles (Figure 3c). Herein, we refer to true protrusion simply as protrusion.

The topographic identification of layers and the creation of protrusion profiles begins by extracting elevation profiles near-perpendicular ($<20^\circ$ relative layer orientation) to the layer orientation from the HiRISE DTM. Error propagation in protrusion measurements due manual drawing of elevation profiles was assessed through elevation profiles taken between perpendicular (0°) to 45° relative to layer orientation. We found that error propagation in protrusion measurements from non-perpendicular profiles was <0.5 m. At each site, elevation profiles were extracted in sets of 5, with each set of profiles separated by ~ 100 m across the trough walls from east to west (Figure 2). Each profile within a set is separated by 12.5 m to obtain an average of the surface expression over 100 m horizontal distance, thus reducing the noise caused by roughness along strike making it easier to identify MB peak protrusion and the recessive points bounding the MB (a similar approach to reducing this noise and obtaining average protrusion profiles was used by Becerra et al. 2016). Herein, we refer to these bounding strata/topographically recessive layers as “interbeds” since they reside between the protruding MBs. At N0, 38 sets of profiles were extracted over ~ 6.5 km along the trough wall from east to west. At N8, 33 sets of profiles were extracted over ~ 8.2 km along the trough wall from east to west.

Protrusion profiles are produced by detrending or normalizing the elevation profiles using methods established by Becerra et al. (2016, 2019) (Figure 3). These profiles allow us to confidently identify the two upper and lower recessive points of the interbeds, and the peaks of MB protrusions, which are otherwise a challenge to locate in the elevation profiles (Figure 3b vs. 3c). The detrending code uses a moving window of a set size, which means protrusion profiles are normalized to the local slope defined by the window size.

Thus, this topographic detrending acts as a high-pass filter, suppressing wavelengths larger than the window size (Becerra et al., 2016, 2019). The smaller the window, the easier it is to resolve and identify thinner layers. Larger windows highlight thicker layers and make it more difficult to resolve thinner ones. Therefore, we chose a window size of 250 to resolve the thinner layers higher up in the stratigraphic sequence and 350 to resolve the thicker layers lower in the sequence. Since the protrusion profiles have been normalized by standard local slope windows, they are not true protrusion values. Once the elevation of protrusion peak and the recessive points have been obtained from the protrusion profiles, the net topographic relief of a MB can be measured from the original elevation profiles (i.e., true protrusion). Subset Figure 3d shows how true protrusion (distance between P and P') is calculated using the elevation points of the MB protrusion peak and interbed recesses.

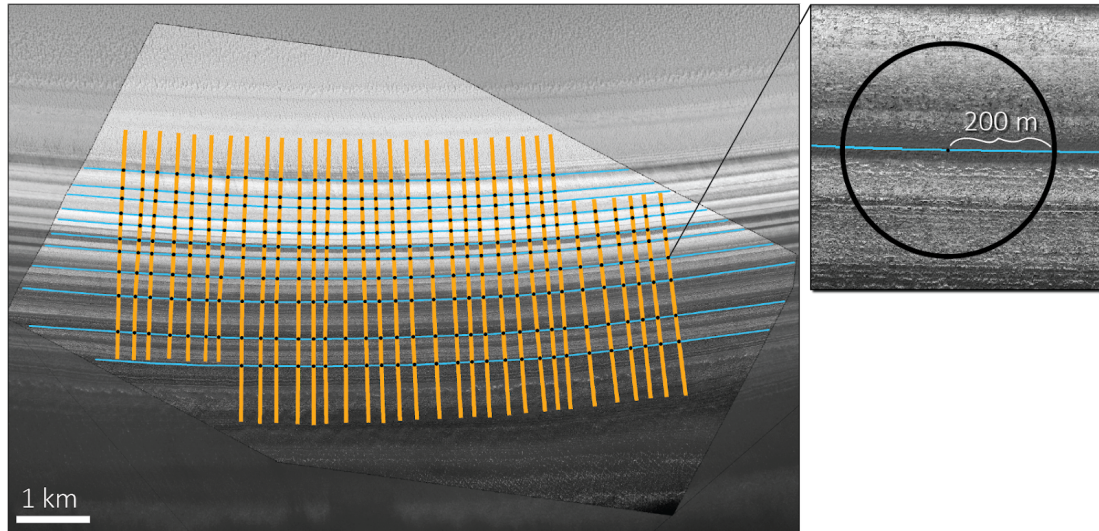


Figure 2. Example of elevation profile sets (orange) taken along trough site N8. Each orange line, which looks like a single thick profile is actually 5 profiles but since they are so close together, they cannot be individually resolved in this figure. Blue lines trace MBs 8, 6, 5, 3, 2, 1, 0, -1, and -2 in order from top to bottom. Black dots are the intersection points of individual MBs and the middle of each profile set (5 profiles per set). The subset image to the right shows how the intersection points are used to obtain the average brightness and slope within a 200 m radius around the center of a set of profiles.

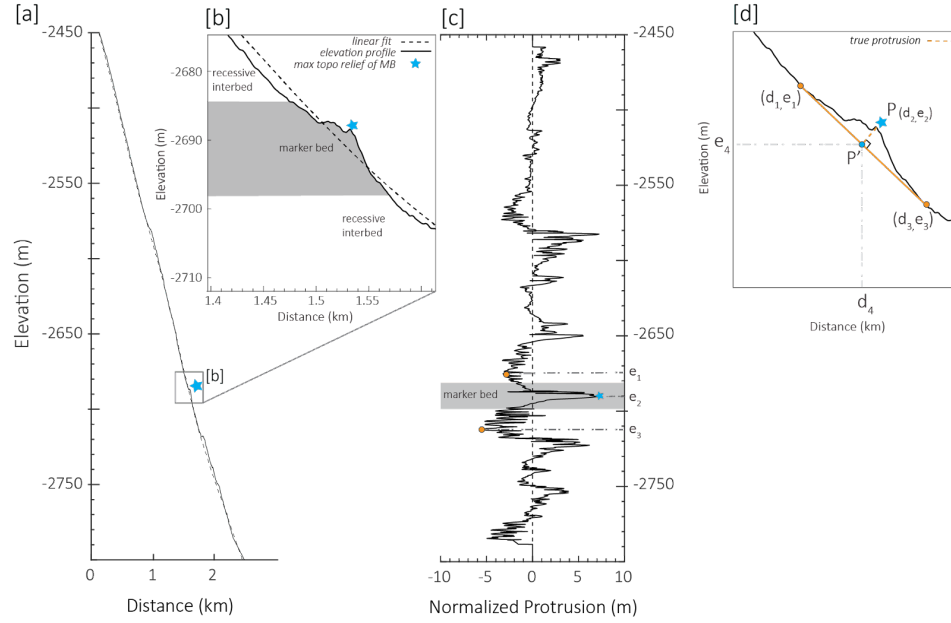


Figure 3. Steps used to calculate the true protrusion of an individual marker bed from elevation profiles. The average elevation profile [a] from a profile set is turned into a depth-dependent protrusion profile [c] by normalizing/detrending the elevation profile via a local linear fit. A local linear fit window [b] is moved down the elevation profile to produce the linear fit that is adjusted to the local trough slope allowing for the layers to become visibly prominent and easily identifiable as shown in the normalized protrusion profile. Once the protrusion profile is made, the elevation points of the protrusion peak (blue star) and elevation points of the upper and lower interbed minima are identified (orange circles). These three elevation points are taken back to the elevation profile to calculate the distance that a MB protrudes relative to IB recession and normal to the slope of a continuum connecting the two recessive points as illustrated in [d].

2.1.3 Statistical correlation

Large-scale topographic trends drive katabatic wind speed (100's of meters to few kilometers) and control the amount of incident solar radiation, while dust cover controls the absorption of solar radiation through its albedo and thermal properties. In turn, the deposition and distribution of dust is controlled by slope and wind speed. To assess whether protrusion variability for a given MB is due primarily to the large-scale topography or to the dust cover, we performed a Bayesian Kendall correlation analysis for protrusion versus slope and protrusion versus brightness (brightness as proxy for dust veneer cover). For this statistical analysis, we obtained the average brightness and slope (from the 100 m/pixel slope map) within a 200 m radius around each MB-profile set intersection using the zonal statistics tool in ArcMap (Figure 2). A radius of 200 m (diameter of 400 m) was chosen to remain on the same order of magnitude length scale for the horizontal resolution of grids used in mesoscale modeling of slope winds (e.g., katabatic winds) (e.g., Spiga and Smith, 2018). Additionally, radius of 200 m covers both the MB and partial to full coverage of bounding

interbeds for all the MBs (regardless of MB ground-thickness). Since the IBs are brighter when not blanketed by dust we used the lateral change in combined average brightness (i.e., driven primarily by dust cover) of each MB and the IBs above and below it as a proxy for change in dust cover on the wall.

The Kendall rank correlation coefficient, τ_B , measures the association or dependency between two measured quantities. Kendall correlation analysis assumes a monotonic relationship between each pair (not a linear relationship) and does not assume the data are normally distributed. These assumptions are appropriate for our data because would data follow these assumptions when plotted separately and as histograms (not shown). The data from N0 and N8 were combined for each shared MB. Then, the data (protrusion, brightness, and slope) for each MB were normalized between 0 and 1 for the correlation analysis. Normalizing allows for proper statistical comparison and correlation of data values with differing units of measurements and ranges. The correlation analysis outputs a Bayes Factor number and τ_B coefficient value. A high Bayes Factor (>1) means there is evidence for the alternative hypothesis (H1), which means there is evidence for either positive or negative correlation (i.e., there is a dependency between the pair). A low Bayes Factor (<1) means there is evidence for the null hypothesis (H0), which means there is evidence supporting a lack of correlation (either positive or negative) or that the variables are independent. The τ_B coefficient has values between -1 and 1, with each extreme indicating a perfect negative (-1) or positive (+1) dependency, respectively, and 0 indicating no relationship.

2.2 Observational evidence: Results

2.2.1 Brightness and slope versus protrusion

We found that not all MBs behave stratigraphically the same when it comes to lateral variability in protrusion. MBs stratigraphically lower in the section (MB 3, 2, 1, 0, -1, -2, -3) exhibit consistently strong evidence for a positive correlation between protrusion and slope and protrusion and brightness (Table 2). In other words, these MBs protrude more on the steeper, dust veneer-free (i.e., brighter) sections. This is apparent when comparing the slope, brightness, and protrusion patterns between sites N0 and N8 in Figure 4 and Figure 5. N0 shows both a drastic slope and brightness change from east to west (Figure 4). For example, MB1's protrusion in the eastern section has a max of 6.5 m, with a local slope of 19.5°, and an integrated MB-IB brightness of 0.5. Moving west, MB1's protrusion decreases to a minimum of 0.5 m, with a local

minimum slope of 6.5° and brightness of 0.35. At N0, there is a clear relationship between the decrease in slope, the decrease in albedo due to the increased presence of a dust veneer, and the decrease in protrusion.

Unlike N0, N8 does not exhibit strong changes in slope or dust cover laterally ($\Delta\text{slope} < 2^\circ$; $\Delta\text{brightness} < 0.02$). As a result, N8 protrusion also does not exhibit lateral changes. N8 protrusions remain consistent laterally over eight kilometers (average protrusion of 1.5 m with a range of 0.5–3 m). These site-specific patterns help expand on and highlight the MB-specific correlation results in Table 2. MBs with a strong positive correlation between slope and protrusion have an equally strong positive correlation between brightness and protrusion (MBs 3, 2, 1, 0, -1, -2, -3) (Table 2). Our results suggest that the change in protrusion for the older, lower strata is a function of both local trough wall slope and the difference in brightness between the interbeds and MBs; but what exactly are the implications of this in terms of nonuniform sublimation? Observational data alone is insufficient to detangle the effects of persistent dust cover/content from local trough topography (~200–400 m scale) since the dust cover and slope today (more so the dust) was likely different than present-day. The role that these, and more, factors have on nonuniform ice loss is explored through sublimation modeling in Section 3 and 4.

Table 2. Bayesian Kendall's correlation results in stratigraphic order for N0 and N8 data combined

Slope vs Protrusion				Brightness vs Protrusion	
<i>Layer</i>	<i>n</i>	τ_B	BF_{10}	τ_B	BF_{10}
<i>MB8</i>	57	0.03	0.18	-0.195	1.65
<i>MB6</i>	59	0.128	0.46	0.276	18.47*
<i>MB5</i>	63	-0.381	2.38E+03***	-0.231	5.51
<i>MB3</i>	65	0.566	4.63E+08***	0.474	7.05E+05***
<i>MB2</i>	64	0.488	1.36E+06***	0.341	3.98E+02***
<i>MB1</i>	67	0.54	1.30E+08***	0.593	8.32E+09***
<i>MB0</i>	66	0.441	1.15E+05***	0.335	3.83E+02***
<i>MB-1</i>	67	0.485	2.46E+06***	0.538	1.06E+08***
<i>MB-2</i>	59	0.348	2.83E+02***	0.412	5.65E+03***
<i>MB-3</i>	29	0.488	1.78E+02***	0.419	31.43**
Correlation Legend					
* $BF_{10} > 10$ (strong evidence H1), ** $BF_{10} > 30$ (very strong evidence H1), *** $BF_{10} > 100$ (extreme evidence H1) - τ_B = negative correlation + τ_B = positive correlation					

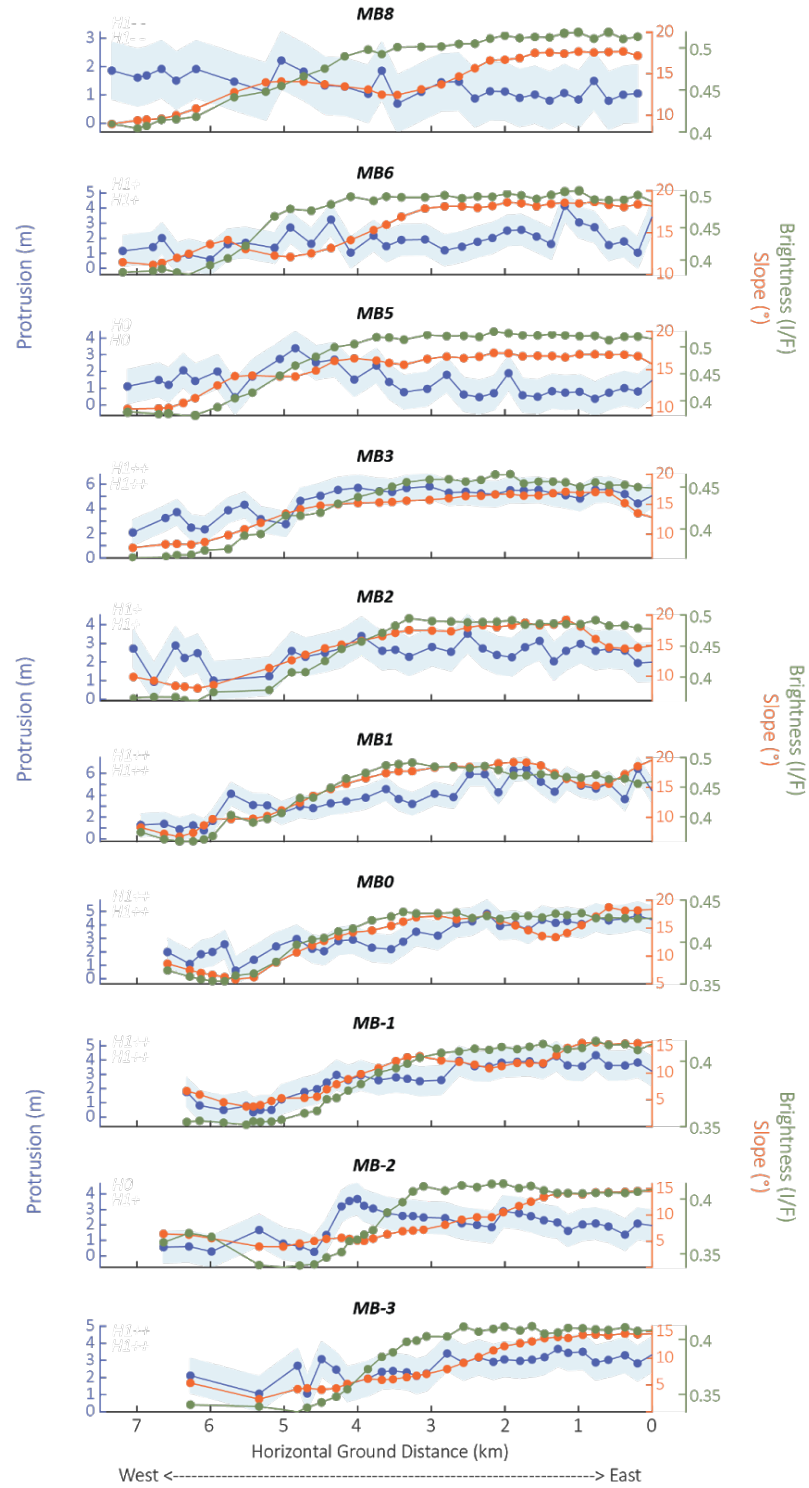


Figure 4. Protrusion (blue), slope (orange), and brightness (green) plotted as a function of horizontal ground distance from east to west for each protruding MB layer at trough site N0. Plots are in stratigraphic order from the top. The slope and brightness are extracted at each layer, and the profile set intersection (each potted orange and green dot) is the average zonal slope and brightness within a 200 m radius of the intersection. This average brightness encompasses the two recessive layers plus the bounding protruding layer.

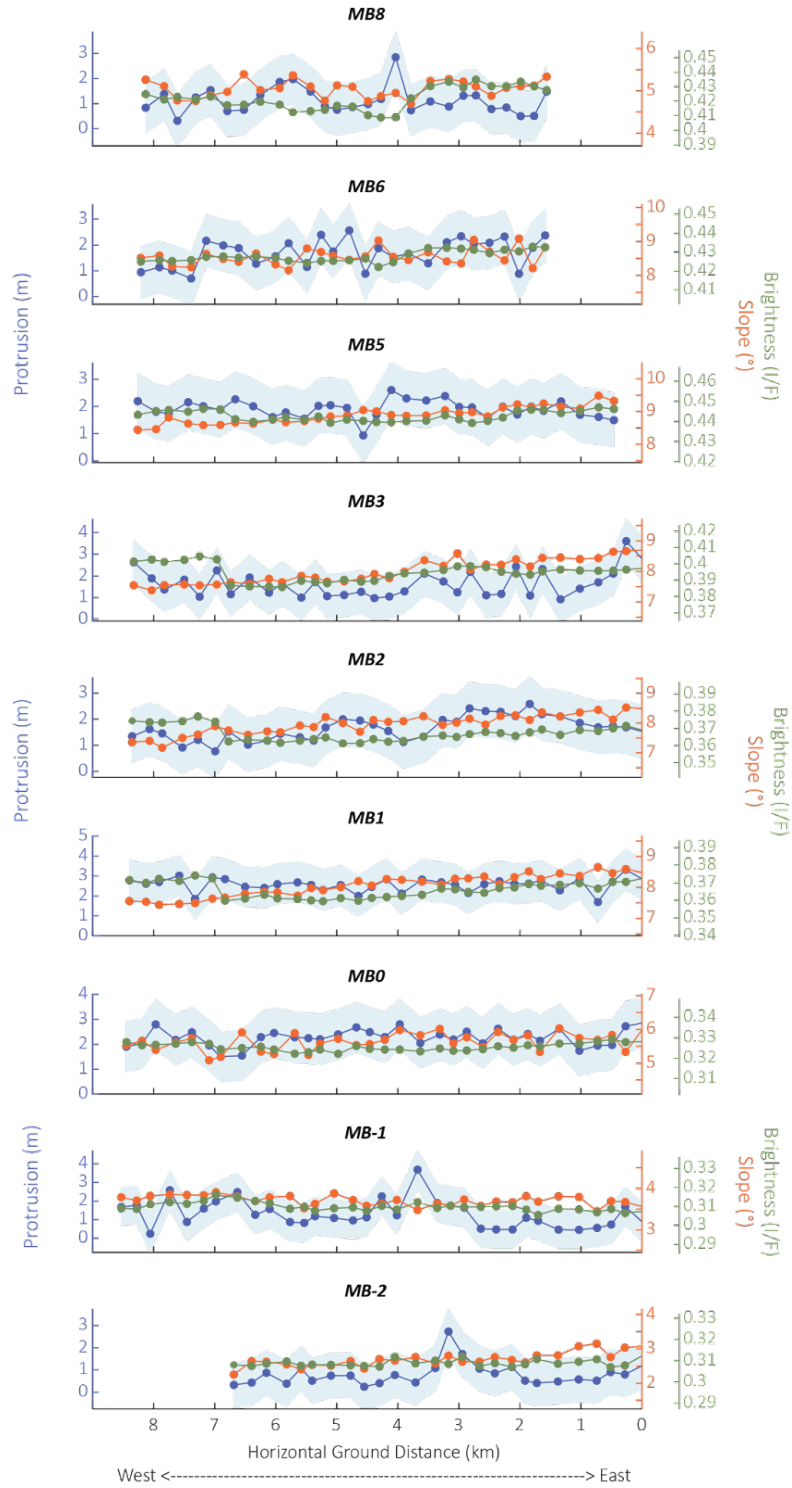


Figure 5. Protrusion (blue), slope (orange), and brightness (green) plotted as a function of horizontal ground distance from east to west for each protruding MB layer at trough site N8. Plots are in stratigraphic order from the top. The slope and brightness are extracted at each layer, and the profile set intersection (each potted orange and green dot) is the average zonal slope and brightness within a 200 m radius of the intersection. This average brightness encompasses the two recessive layers plus the bounding protruding layer.

3 Modeling ice instability

The lateral change in protrusion is a function of the erosional resistance of the MB (controlled by a combination of its dust content and the degree of induration of the dust lag) and the total sublimation difference between the MB and the bounding lighter-toned strata. Wall slope and dust cover (and thus, albedo) contribute to the spatially nonuniform ice sublimation shown in Section 2. However, the effect of persistent dust cover and/or dust content versus local trough topography (~200–400 m scale slope) on sublimation's spatial and temporal variability remains to be disentangled. In addition, when and how long would it take for the lateral difference in the stratum's expression to occur? The goal of using this sublimation model is to constrain the amount of time it takes to develop the spatial variability we see today, particularly at N0, where we see the most spatial variability in protrusion over a few km.

3.1 Sublimation model

3.1.1 Model set-up

We model the ice stability and sublimation processes occurring on the trough wall using the model developed by (Bramson et al., 2017, 2019). Our goal is to test if sublimation differences between different NPLD strata can generate observations of protrusion. The model can include diffusive and both free and forced convective ice loss. Diffusive ice loss occurs following water vapor flux through and out of the subsurface, which follows the diffusion law, where the diffusion coefficient is set by the medium through which the flux occurs (e.g., regolith/dust lag). Convective ice loss occurs following vapor flux with the atmosphere due to sensible heat caused by buoyancy of the near-surface air (free) and advection due to wind (forced). The model description and results presented below are for convective ice loss only since we assume there is no pre-existing dust cover. We tested (see Supplementary Materials S1) for diffusive ice loss through a lag and found that a few millimeters quickly reduces sublimation to 0, so the wall experiences little time undergoing diffusive loss compared to the time spent either stable to diffusive loss unstable due to convective loss.

3.1.2 Thermal conduction model

The model uses a 1D semi-implicit thermal conduction model that simulates surface energy balance and transfer of heat between subsurface numerical layers, applicable to surface ice with no lag cover, using Equations 1 and 2. Surface temperatures (T_{surf}) and the transfer of heat through the subsurface are calculated, respectively, at a timestep of 1500 seconds throughout a martian year following

$$\varepsilon \sigma T_{surf}^4 = S_o \cos i (1 - a) + L_{CO2} \frac{dm_{CO2}}{dt} + k \frac{\partial T}{\partial z} + IR_{\downarrow} \quad \text{Eqn. 1}$$

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(k \frac{\partial T}{\partial z} \right) \quad \text{Eqn. 2}$$

The left-hand side of Equation 1 is the amount of energy radiated by the surface for a given T_{surf} , and ε and σ are the emissivity of the surface and the Stefan Boltzmann constant, respectively. The first term in the right-hand side of Equation 1 is the amount of solar energy absorbed by the surface, where S_o is the solar flux at Mars including an additional 2% top-of-atmosphere flux due to scattered light in the atmosphere (Kieffer et al., 1977; Aharonson and Schorghofer, 2006; Bramson et al., 2019), i is the incidence angle (angle of the sun from zenith relative to surface slope) which is influenced by surface slope, aspect, and latitude, and a is the prescribed surface bond albedo (see Table 4). The second term on the right-hand side accounts for the presence of CO₂ frost, when applicable. L_{CO2} is the latent heat of CO₂ (590 kJ/kg), and m_{CO2} is the mass of CO₂ frost on the surface calculated at every timestep. When T_{surf} falls below the frost point of CO₂ for 610 Pa atmospheric pressure, T_{surf} is then fixed to the CO₂ frost point. The third term on the right-hand side accounts for the exchange of heat with the subsurface. Lastly, the fourth term on the right-hand side provides additional energy from downwelling infrared radiation, modeled as 4% of the daily noontime flux (Aharonson and Schorghofer, 2006). Heat conduction (Equation 2 and the third term in the RHS of Equation 1) depends on the thermophysical properties (thermal conductivity (k), density (ρ), and heat capacity (c)) of each numerical layer in the model. k is the thermal conductivity of a given modeled thermophysical layer, which we calculated assuming a linear volumetric mixture of ice and rock (Siegler et al., 2012). The thermophysical properties of endmembers assumed for rock and ice can be found in Bramson et al. (2017) and Table 3. The thermophysical properties used to simulate our strata can be found in the same table.

3.1.3 Ice loss calculations

We use our modeled temperatures from the thermal conduction model to calculate the amount of sublimation from the ice during periods of ice instability. In the case of buried ice (i.e., covered by a lag deposit), the saturation water vapor density at the ice interface (ρ_{vapor}) is calculated using the modeled temperatures and the Clausius-Clapeyron equation and ideal gas law. Ice is stable if $\rho_{vapor} \leq \rho_{atmos}$. If $\rho_{vapor} > \rho_{atmos}$, then ice is not stable and will sublimate until $\rho_{vapor} = \rho_{atmos}$. In cases of exposed surface ice, the amount (mass flux) that would sublimate at each timestep is calculated using the equations for free and forced convection following Dundas and Byrne (2010) see Equations 3–5 (and previously utilized in Bramson et al., 2019).

$$m_{forced} = M_w \frac{1}{KT_{bl}} A u_{wind} (e_{sat} - e) \quad \text{Eqn. 3}$$

$$m_{free} = 0.14 \Delta \eta \rho_{ave} D \left(\left(\frac{\Delta \rho}{\rho} \right) \left(\frac{g}{v^2} \right) \left(\frac{v}{D_{at}} \right) \right)^{\frac{1}{3}} \quad \text{Eqn. 4}$$

$$m_{sub} = m_{forced} + m_{free} \quad \text{Eqn. 5}$$

The forced convection equation (Equation 3) models the effect of water vapor transport by winds on the sublimation process. This equation does not model the effects of wind on mechanical erosion. In the forced convection equation, u_{wind} is the average annual wind speed, and A is the drag coefficient for an ice surface. We test three different wind speeds of 0, 10, and 20 m/s, using 10 m/s as the nominal wind speed. These wind speeds were chosen based on mesoscale modeling of katabatic winds down spiral trough walls from Spiga and Smith (2018). The drag coefficient for an ice surface is set to 0.002 (Dundas and Byrne, 2010; Bramson et al., 2019). The saturation vapor pressure of water in the atmosphere and the surface are e and e_{sat} , respectively. M_w is the molecular mass of water, and T_{bl} is the temperature of the boundary layer, which is set to be the average between the ice surface temperature and the atmospheric temperature (Dundas and Byrne, 2010; Bramson et al., 2019). The nominal atmospheric water vapor pressure (e , or ρ_{atmos} when converted to density) is calculated based on the orbital configuration at every timestep from the present going back to 300 kyr. See Bramson et al. (2019) for further details. Similar to our range of wind speeds, we run the model for a range of atmospheric water vapor partial pressures/densities (0 and nominal).

Table 3. Parameters used in the thermal model and convective ice loss calculations of Equations 1–5

Variable	Parameter	Comment
ε	emissivity of the surface	0.8 if covered by CO ₂ frost, 1 otherwise
σ	Stefan-Boltzmann constant	$5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$
T_{surf}	surface temperatures	calculated at every time step
S_o	solar flux of Mars	calculated at every time step through a martian year using Laskar (2004) orbital solutions
i	incidence angle of the sun	calculated based on latitude and slope of terrain
a	bond albedo	see Table 5
L_{CO_2}	latent heat of CO ₂ frost on the surface	590 kJ kg^{-1}
m_{CO_2}	mass of a CO ₂ layer on surface	dependent on latitude and time of year. Calculated at every time step.
k	thermal conductivity	See Table 4
$IR_{\downarrow}$	downwelling infrared from the atmosphere	assumed to be 4% of noontime insolation
I	thermal inertia	See Table 4
ρ	density	See Table 4
c	heat capacity	See Table 4
M_w	molecular mass of water	$18.01528 \text{ g mol}^{-1}$
T_{bl}	temperature of the boundary layer	average T_{surf} and T_{atm} ; $T_{\text{atm}} = T_{\text{min}}^b T_{\text{reg}}^{1-b}$. $b=0.2$ following (Dundas & Byrne, 2010)
A	drag coefficient	0.002 for an icy surface (Dundas & Byrne, 2010)
u_{wind}	wind speed	See Table 5
e_{sat}	saturation vapor pressure	calculated for each T_{surf} using C-C equation
e	atmospheric water vapor partial pressure	calculated at every time step
K	Boltzmann constant	$1.38064852 \times 10^{-23} \text{ m}^2 \text{ kg s}^{-2} \text{ K}^{-1}$
ν	kinematic viscosity	calculated following Dundas & Byrne (2010)
g	gravity	3.721 m s^{-2}
D_{at}	diffusion coefficient of H ₂ O in CO ₂	calculated following Dundas & Byrne (2010)

In the free convection equation (Equation 4), $\Delta\eta$ is the dimensionless difference between atmospheric and surface gas water mass fraction, which is calculated in the same manner as Dundas and Byrne (2010) and Bramson et al. (2019). D is the diffusion coefficient of H_2O in CO_2 , g is gravity, ν is the kinematic viscosity, and $\Delta\rho/\rho$ is the difference between the atmospheric gas density and that of the surface divided by a reference density (Dundas and Byrne, 2010; Bramson et al., 2019). Total sublimation from free and forced convective terms are summed together. The mass fluxes calculated from Equation 5 are integrated over 1 year and converted to the rate of ice sublimation in mm/year, perpendicular to the sloped trough wall.

3.1.4 Strata properties

We calculate the sublimation difference between two compositionally distinct strata, marker beds (MB) and interbeds (IB). As suggested by their lower apparent brightness, the darker protruding MBs must contain more internal dust, while the interbeds are typically lighter-toned and thus, contain less internal dust. In addition, the HiRISE protrusion profiles show interbeds being recessed compared to the protruding MBs. We use the protrusion measurements of MB0 at site N0 (Section 2.2.) as our benchmark stratum to compare the model results. The thermophysical properties of a modeled MB and IB can be found in Table 3. The composition and albedo used for modeled strata can be found in Table 4, and the number of strata scenarios is illustrated in Figure 6.

To understand the processes that cause strata protrusion, we model and compare the sublimation rates between the MB and IB. The MB stratum is modeled with 30% dust with two different surface albedos (0.15 and 0.25). This range in bond albedo is modeled to fully bracket the low albedos of troughs previously observed (Kieffer et al., 1976; Kieffer, 1990; Paige et al., 1994). The dust abundance was chosen based on radar-derived average dust abundance of individual subsurface radar reflectors—believed to be but have yet to be genetically linked to MBs—near location N0 (Lalich et al., 2019). The interbed is modeled at three different albedos (0.35, 0.25, and 0.15), but all cases are modeled with internal dust content of 3% dust. The dust abundance is consistent with the estimated bulk composition of the NPLD (Grima et al., 2009). An albedo of 0.35 represents a bright, bare, dust lag/veneer free interbed (IB1) as it is the highest trough albedo observed (Kieffer et al., 1976; Kieffer, 1990). To assess the effect of a thermally thin dust cover on the ice-rich interbed (<a couple mm's; not enough dust to insulate the ice below it), we modeled the second type of

interbed, IB2, with an albedo of 0.15 and 0.25. Preliminary tests (Supplementary Material, Figure S2) of modeling sublimation through a lag of 2 mm showed the presence of a thin lag has a negligible effect on peak sublimation rates when compared to a no-lag scenario (where diffusion through a sublimation lag is not modeled). We found that lag thicknesses above 2 mm had a substantial effect on the sublimation process. Sublimation effectively shuts down with lag over 5 mm thick (Supplementary Material, Figure S2). Due to these results, we run IB2 under the assumption that low surface albedo is induced by a very thin dust cover where diffusion has a negligible effect and thus is not modeled.

The modeled outputs of annual lag production rate and sublimation rate provided here are instantaneous for bare (0 mm lag) strata at each 1000-year interval of the orbital solution (Laskar et al., 2004), rather than accounting for secular changes to the lag thickness (caused by lag production and erosion processes). If the lag were allowed to grow—thus accounting for the lag developed in the previous time step without modeling mechanical erosion the lag—it would eventually reach a thickness that would cause the sublimation of ice to be inhibited or completely shut down. Section 4 provides an example framework of secular changes if lag build-up and erosion were considered.

To understand the cause of the lateral variability in protrusion, we model ice stability between the far east and west wall locations at N0 for a MB versus an IB. In total there are six scenarios (Figure 6), three for each (western and eastern) location using their respective aspects and slope. The slope and aspect for the east and west locations used in the modeling were the averages of each from a profile spanning across MB0 and its adjacent interbeds. The values were extracted from the downsampled HiRISE DTM slope and aspect maps (100 m/pixel) in order to obtain slope and aspect at the local trough scale and avoid the smaller-scale roughness associated with layer texture captured by a 1 m/pixel high-resolution maps. The westernmost portion of the trough wall at site N0 faces nearly due east at a slope of 6°, while the eastern portion faces southeast at a slope of 11°.

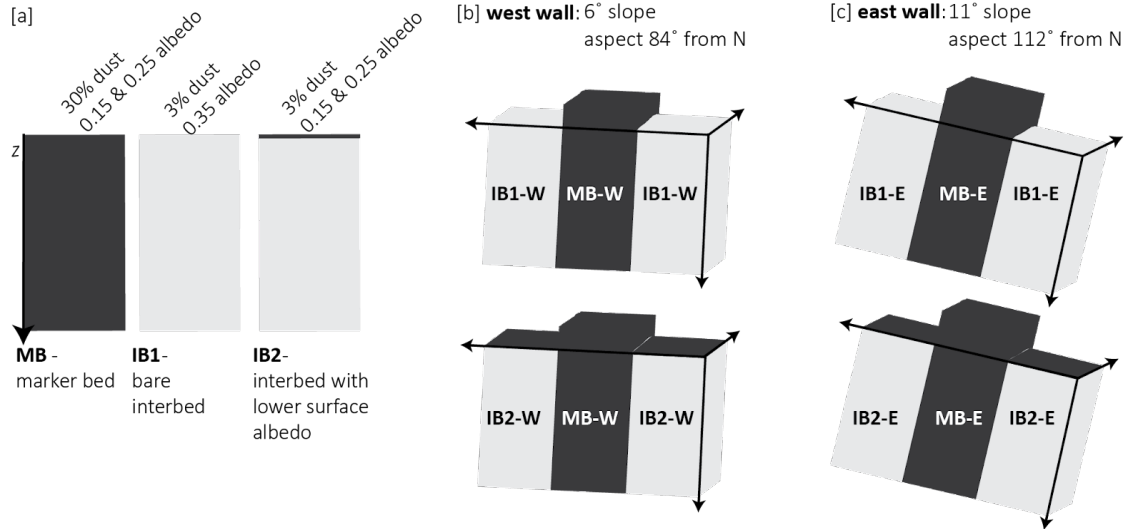


Figure 6. Illustration of the model set-up. The model was run for 3 different stratum exposure scenarios at the slope and aspect of the west and east wall. Marker beds (MB) and interbeds (IB) are modeled with different ice-dust ratios (30% and 3% dust, respectively). For each stratum scenario at either wall site, the water vapor density and wind speeds were set at 0 and nominal (and 2x nominal for wind speed) (see Table 4). The MB and IB2 strata were modeled at two different albedos to cover the expected albedo range of an average dusty trough wall (wind speed and water vapor assumed nominal). As a result, our parameter space resulted in modeling a total of 28 runs (MB and IB2 = 5 runs; IB1 = 4 runs).

Table 4. List of thermal inertias and density*heat-capacities for the material endmembers (for reference) and for the four thermophysical cases used in the model. These are the same properties used in Bramson et al. (2019) except for the addition of the marker bed properties

Material	Thermal Inertia ($\text{J m}^{-2} \text{K}^{-1} \text{s}^{-0.5}$)	ρc ($\text{J K}^{-1} \text{m}^{-3}$)
Marker bed (70% ice, 30% dust)	2195	1,820,390
Interbed (97% ice, 3% dust)	2,147	1,457,159
Endmember rock ($k=2 \text{ W m}^{-1} \text{K}^{-1}$, $\rho=3300 \text{ kg m}^{-3}$, $c=837 \text{ J kg}^{-1} \text{K}^{-1}$)	2,350	2,762,100
Endmember ice ($k=3.2 \text{ W m}^{-1} \text{K}^{-1}$, $\rho=920 \text{ kg m}^{-3}$, $c=1540 \text{ J kg}^{-1} \text{K}^{-1}$)	2,129	1,416,800

Table 5. Primary free parameter ranges used in thermal and ice sublimation calculations

Parameter	Values (*denotes nominal)	Comment
<i>Atmospheric water vapor partial pressure (e, or ρ_{atmos} when converted to density)</i>	0, nominal	Dundas et al. (2015); nominal values are calculated at each time step based on the orbital solution.
<i>Wind speed (m/s)</i>	0, 10*, 20	Range of values based on Smith and Spiga (2018).
<i>Volumetric dust content</i>	3% ^a , 30% ^b	^a Interbed dust abundance based on bulk NPLD composition from Grima et al. (2009), ^b Marker bed abundance based on average estimates from Lalich et al. (2019)
<i>Bond albedo</i>	0.15, 0.25, 0.35	Lowest albedo is used for both marker bed and interbed-2. Interbed-2 was also modeled at 0.25 (average trough wall albedo). Interbed-1 is modeled at 0.35 to reflect the albedo of average bare interbed ice. 0.25 albedo runs assumed nominal wind speed and water vapor partial pressure.
<i>Slope</i>	6 ^c , 11 ^d	Average slope calculated from HiRISE DTM elevation profile of case study layer MB0 at trough site N0 (perpendicular to layer orientation); degrees from horizontal. ^c West wall. ^d East wall.
<i>Aspect/trough orientation (°)</i>	84 ^c , 112 ^d	Same location that slope profile was taken (from downsampled HiRISE); degrees clockwise from north. ^c West wall. ^d East wall.

3.2 Model Results

Under present-day conditions, the ice is generally stable unless certain conditions are met, such as high wind speeds, a low albedo surface, or the absence of water vapor. In the last several hundred thousand years, the sublimation and subsequent lag production of our model strata were subjected to 3 prolonged periods of ice instability (Figures 8 and 9). The first occurred between 60–125 kya, corresponding to two solar insolation peaks at 80 and 110 kya. This period corresponds to a peak in eccentricity ~75 kya and one in obliquity ~100 kya, with perihelion occurring during northern summer. A second, smaller single instability period occurred between 150–180 kya, peaking at 165 kya. Lastly, the strongest instability period of the three occurred between 220–300 kya peaking at 275 kya. These peaks mainly coincide with the season of perihelion occurring in northern hemispheric summer. These instability peaks also coincide with, and are

amplified by, small peaks in either obliquity ($\sim 27^\circ$) and or large peaks in eccentricity affecting the polar insolation.

3.2.1 Effect of albedo and dust content of strata on sublimation and lag production rates

MB protrusion is the total sublimation difference between the MB and the IBs that sandwich it. We find that MB sublimation rates are higher than bare ice-rich interbeds (IB1) due to the strong dependency of albedo on temperatures and thus sublimation rate. We found that an IB1 (albedo of 0.35) remains stable and does not undergo sublimation unless under higher insolation conditions at 270 kya or when atmospheric water vapor is extremely low. The only time the interbed sublimates outside of those two conditions is when the stratum has a surface albedo lower than 0.35 (0.25 or 0.15). Lower albedo for an ice-rich interbed (i.e., IB2) can occur if there is a thin mantle of dust on the surface of the ice-rich interbed, thus lowering the surface albedo allowing for sublimation of the ice beneath it (Figure 7).

Additionally, spectral calculations suggest that even extremely small ($<1\%$) dust content in ice can significantly reduce the albedo (Khuller and Christensen, 2021). For example, at 80 kya, if the MB is darker (albedo 0.15) and the IB is slightly brighter (IB2: albedo 0.25), the sublimation rates would be ~ 7.5 mm/yr and ~ 1.5 mm/yr, respectively; or a 6 mm/yr difference in sublimation rates between the two strata. When the MB and the IB have the same albedo, the interbed sublimates at the same rate as the MB. Therefore, to generate differential sublimation between the MB and IB that would possibly generate positive protrusion for the MB and not the IB, the IB must have a lower bond albedo than 0.35, which can be accomplished by a minor amount of lag/dust veneer. Lower bond albedo enables sublimation (similar to what was proven in Bramson et al., 2019 about trough migration). This concept is discussed in Section 4.

Dust lags develop on ice deposits when dirty ice sublimates, leaving behind the dust contained within it. We calculate lag production rate, in mm/yr, which depends on the sublimation rate, dust content of the modeled stratum, and the assumed porosity of the resulting lag (which we assume to be 40%) (Schorghofer, 2010; Bramson et al., 2019). Figure 8 shows that the MB produces an order of magnitude more lag than the interbed; this reflects that the modeled interbed contains 3% dust and the MB contains 30%. For example, the results for the east wall show that the MB would produce roughly ~ 5 mm of lag at 80 kya, while

during that same year, the interbed would only produce ~ 0.5 mm of lag. The importance of the difference in lag production rate is discussed in Sections 4 and 5.

3.2.2 Effect of wind speed and trough topography on sublimation and lag production rates

Compared to albedo, the difference in the sublimation rate induced by the change in surface slope and trough orientation from the east versus west wall is small. For example, at 80 kya (and assuming slope and aspect where the same then as are presently), the nominal model runs (albedo 0.15) found a difference of 3 mm/yr in sublimation and 2 mm/yr in lag production for the MB between the eastern wall and the western wall (the east wall has a steeper slope of 11° and is facing southeast; the west wall has a shallower slope at 6° with the wall facing due east).

Wind speed has a significant effect on the sublimation and lag production rate (Figure 7 and 8). In Figures 8 and 9, the lower boundary of the light gray shaded area is the model output for an average annual wind speed of 0 m/s, and the upper boundary of the shaded area corresponds to a wind speed of 20 m/s. Therefore, wind speed has a similar effect on sublimation and the lag production rates as the albedo of the surface. For example, a wind speed of 0 m/s with an albedo of 0.15 results yield very similar sublimation rates to the model run with a nominal wind speed of 10 m/s but an albedo of 0.25.

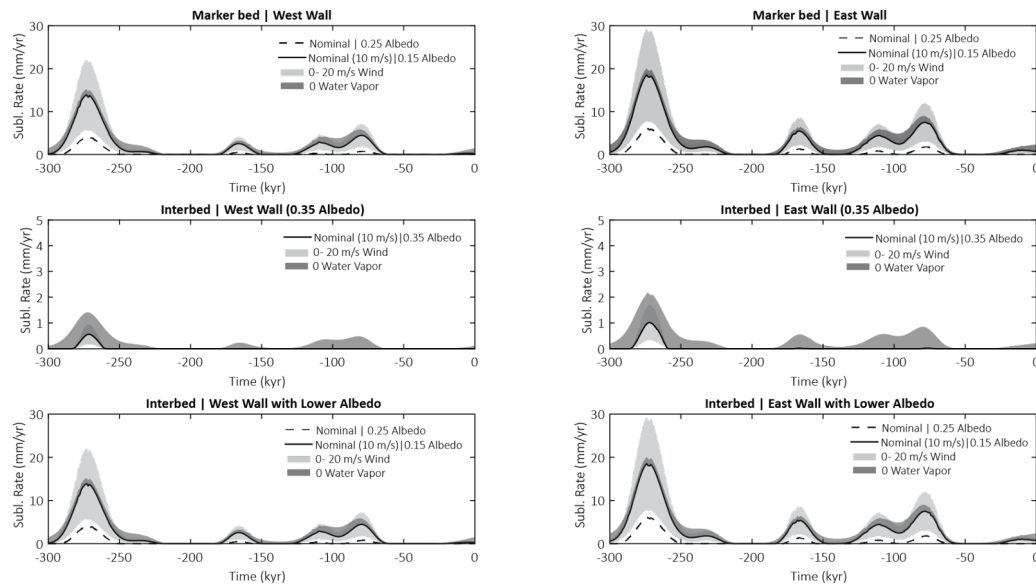


Figure 7. Instantaneous sublimation rate (mm/yr) plotted as a function of time for the various layer scenarios and model parameter space (strata albedo, wind speed, water vapor, trough slope and orientation). Left and right plots show the model outputs for the west and east wall slope and orientation, respectively. The top-down order of the plots is MB, IB1 and IB2.

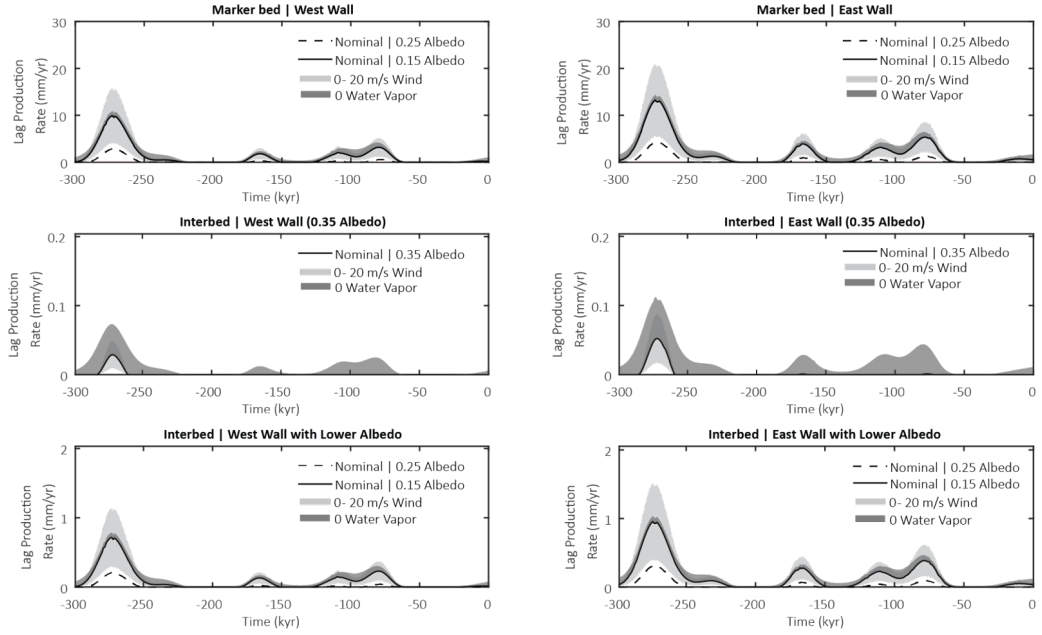


Figure 8. Instantaneous lag production rate (mm/yr) for all runs plotted as a function of time for the various layer scenarios and model parameter space (strata albedo, wind speed, water vapor, trough slope and orientation). Left and right plots show the model outputs for the west and east wall slope and orientation, respectively. The top-down order of the plots is MB, IB1 and IB2.

4 Framework for forming protruding marker beds

The sublimation modeling results presented in Section 3.2 suggest a bare, darker MB will always sublimate more relative to a bare, brighter interbed (IB1) due to the strong dependency of albedo on the sublimation rate. This could be confounding, as it would cause the MB to be the one recessed relative to the interbed, which would be the opposite of what is observed (Section 2 of this work; Fishbaugh et al., 2010; Becerra et al., 2016). If the IB and MB have different albedos—0.25 and 0.15, respectively—the MB would still be the stratum recessed into the wall. If the interbed (IB2) has the same albedo as the MB (e.g., has a negligibly diffusive thin veneer or high enough volumetric fraction of dust to darken the strata), then both strata would sublimate at the same rate, which still would not generate positive protrusion of the MB. However, we find that the key component for producing MBs with positive protrusion is the modulation of lag production versus physical erosion of the lag and the resulting integrated amount of time each stratum spends stable versus sublimating and producing lag.

We put forth a framework for generating marker beds protrusion using the results from the sublimation modeling in Section 3.2 with a constant lag removal rate as a proxy for mechanical erosion independent from erosional resistance. We look at the cyclical nature of sublimation and corresponding lag production and add an assumed constant lag removal rate to obtain net lag growth per year. Herein we describe the processes of marker bed protrusion development when lag removal is considered. During periods of high insolation, the bare trough wall experiences a burst of sublimation. For dustier layers (e.g., MBs in this case), this leads to the generation of a thermally thick lag (in some cases, this can occur within a single summer). This lag is thick enough to inhibit further sublimation of the MB the following years. During this time, the MB only experiences annual erosion of its lag until it is thin enough or completely removed, allowing the MB to experience another pulse of sublimation. While the MB sublimation is shut down, the ice-rich IB continues to sublimate. Due to the IB's dust-poor nature, a very small amount of lag is produced compared to the MB. The IB, therefore, experiences more frequent cycles of sublimation, lag production, and lag removal than the MB, thus allowing the IB to retreat further while the sublimation of the MB remains inhibited. This differential sublimation cycle continues until the MB's lag has eroded to be thin enough to experience another burst of sublimation. We compare the amount of IB ice loss that occurs in each pulse and the frequency that those pulses happen within a single MB cycle (the period of time that the MB's sublimation is shut down due to thick lag) to calculate the net protrusion of the MB.

In the section below, we provide first-order estimates for the time it would take to produce the present-day protrusions (as measured in Section 2.2) using the sublimation rate and lag production rate outputs from the sublimation modeling in Section 3.2. For these first-order estimates, we used the sublimation rate and lag production rates from the two most recent peaks of prolonged polar ice instability periods at 80 kya and 110 kya to see if it is possible to produce this amount of protrusion through differential sublimation conditions of the strata. Here, we use MB0 at site N0 as a case study reference layer, which has an average protrusion of 3 m, with a maximum amount of protrusion of ~5 m in the eastern section of the site and ~1.5 m minimum in the western section. For lag removal we assume a constant rate of 0.05 mm/yr, 0.1 mm/yr, and 0.15 mm/yr. A removal rate of 0.05 mm/yr is the equivalent to a thickness equal to the diameter of a small, clay-sized dust aggregate (Vaughan, 2010) is removed each year. Removal rates of 0.1 and 0.15 mm/yr are the equivalent of removing a single layer of very fine to fine sand-sized aggregates. The goal here is to

iterate the sublimation rate, lag production, and lag removal to estimate how long it takes to produce a protruding MB that matches the average protrusion amount measured in Section 2.2 for MB0.

4.1 Number of years to produce observed MB protrusion?

Table 6 shows the results from our differential sublimation cycle calculations estimating the number of years it would take for a MB to protrude 1.5 m at the west wall and 5 m at the east wall (assuming a constant sublimation and lag production rates at 80 kya and 110 kya from the nominal model runs). A MB can easily reach the observed topographic relief solely from differential sublimation modulated by lag production and lag removal within the most recent prolonged polar ice instability period between 60–125 kyr (within 1,000–25,000 years). The sublimation and lag production rates 80 kya are modeled to be slightly more than the rates 110 kya. Therefore, calculating how long it would take for a MB to reach the observed protrusion using the rate at these two peaks gives us a modest time range estimate.

Under nominal modeling conditions for a range of albedo pairings for the IB and MB and a constant lag removal rate of 0.05 mm/yr, a MB could take roughly ~2,000–5,000 years to reach a protrusion of 1.5 m along the west wall 80 kya (Table 6). 110 kya, it would have taken the marker bed anywhere from ~2,000–15,000 years (Table 6). Under nominal modeling conditions for a range of albedo pairings for the IB and MB and a constant lag removal rate of 0.05 mm/yr, a MB could take roughly ~7,000–10,000 years to reach a protrusion of 5 m along the east wall 80 kya (Table 6). 110 kya, it would have taken anywhere from ~7,000–15,000 years (Table 6). Lastly, the time needed for protrusion formation is the fastest when both the MB and IB are the lowest albedo. The number of years increases by the thousands if the MB and IB are two different albedos or if the MB and IB are the same albedo but brighter.

Table 7 shows the estimated number of years it would take for a MB to protrude under a range of lag removal rates (0.05, 0.1, 0.15, 0.5 mm/yr) and expands beyond looking at the nominal model run conditions to see how different wind speeds affect the time required to reach the measured protrusion. The estimated number of years (Table 7) for assuming annual average wind speeds of 0 m/s and 10 m/s wind for the west wall since are likely more realistic than 20 m/s due to the west wall having an overall lower slope than the east wall and therefore is the more likely to experience slower katabatic wind speeds—assuming the slope difference between the east and west wall section hasn’t changed much of the past 10’s of thousands of years. Between the different wind speed runs, the number of years either increases or decreases by a few

hundred to 1,000 years (Table 7). As expected, the number of years increases when using the rates from the 0 m/s wind speed model run while the number of years decreases when using the rates from the 20 m/s wind speed model run.

The annual lag removal rate has either a positive or negative effect on the number of years. For some scenarios, as the lag removal rate increases, so does the amount of time needed to form the protrusion. Conversely, there are some scenarios where the amount of time decreases. Lastly, some scenarios show the number of years decreasing from 0.05 to 0.1 mm/yr lag removal but reverses (number of years increases) once the lag removal is increased to 0.15 mm/yr. This occurs because lag removal rate of 0.15 mm/yr is greater than IB lag production rate in those specific cases.

Based on the results presented in both Table 6 and Table 7 and comparing them to the rates that were used in the calculations (Supplementary Material Tables S1 and S2), we found two important relationships between the MB and IB layers that increases the number of years needed to produce the amount the protrusion. In both relationships, both conditions (i.e., sublimation rate plus the lag production rate-to-lag removal rate condition) must be met for the time estimate to increase. The two relationships are: 1) if the sublimation rate for the IB is an order of magnitude less than the MB sublimation rate and the IB lag production rate is $\leq$ the lag removal rate, or 2) if both the MB and IB sublimation rates are well under 1 mm/yr and the IB lag production rate is $\leq$ the lag removal rate. However, regardless of these conditions and the factors (dust content, albedo, wind speed, trough slope and aspect) that govern them, the MB topographic signature we see can easily be formed within the most recent peak instability period (~60–125 kya at latitudes ~80–90°N). In the following section, we will discuss what conditions drive MB protrusion, and thus net ice loss of the MB and bounding IBs, to vary over a few km—based on the calculations presented here—and the implication for nonuniform ice loss within spiral troughs cap-wide.

Table 6. The number of years needed to form MB protrusion matching the observed protrusion—MB0 as case study reference: 1.5 m for the west wall and 5 m for the east wall across our albedo parameter space using modeled sublimation and lag rates at 80 kya and 110 kya (from nominal model runs).

rates pulled from	80 kya						110 kya					
albedo pairings	MB & IB 0.15		MB 0.15 & IB 0.25		MB & IB 0.25		MB & IB 0.15		MB 0.15 & IB 0.25		MB & IB 0.25	
location	W	E	W	E	W	E	W	E	W	E	W	E
years (lag removal 0.05 mm/yr)	2,200	7,350	4,890	10,060	4,730	10,025	2,210	7,325	14,735	15,070	12,000	14,580

Table 7. Number of years needed to form MB protrusion with a range of lag removal rates using modeled sublimation and lag rates at 80 kya (includes min and max wind speed runs). *denotes that the MB was never able to reach positive topographic relief.

location	W					E				
albedo pairings	MB & IB 0.15			MB 0.15 & IB 0.25	MB & IB 0.25	MB & IB 0.15			MB 0.15 & IB 0.25	MB & IB 0.25
wind speed(m/s)	0	10	20	10	10	0	10	20	10	10
years (lag removal 0.05 mm/yr)	2,810	2,200	2,330	4,890	4,730	9,075	7,350	6,355	10,060	10,025
years (lag removal 0.1 mm/yr)	1,920	1,100	1,435	6,290	5,600	5,565	4,480	3,665	7,005	6,860
years (lag removal 0.15 mm/yr)	2,085	1,150	955	8,365	6,665	5,950	2,990	2,280	7,725	7,560
years (lag removal 0.5 mm/yr)	3,335	890	515	*	12,000	4,835	1,600	1,500	22,220	11,760

5 Discussion

5.1 Trough wall evolution and development of protruding marker beds

Outlined in Section 4, we developed a framework for developing protruding MBs solely from differential sublimation. In this discussion, we do not discount the genuine likelihood that an indurated lag/erosionally resistant MB surface plays a role in developing the observed topographic relief. The purpose of this work and discussion is to present an endmember explanation for the variable trough wall ice loss.

Our modeling and theoretical calculations suggest the observed protrusion at sites N0 and N8 could have developed on the order of 1000's of years (<25 kyr) via cyclical bursts of differential sublimation modulated by lag removal and variable lag production (a function of dust content and sublimation rate). For context, this timeframe is well within a single martian precession period, where the season of perihelion occurs in the northern hemispheric summer (~25 kyr; a full precession period is 51 kyr).

We find that both stratum types must be unstable for an exposed MB to have a fighting chance in developing into positive topographic protrusion (in the absence of accounting for erosional resistant layer properties). The last time troughs at this latitude (~80-90°N) would have been able to undergo substantial sublimation and develop their present-day NPLD topographic expression would have been 60–125 kya. This time period is consistent with suggested mass trough erosion events (Rodriguez et al., 2021). Troughs at lower latitudes would experience longer summers and overall high insolation annually—and depending on the strata albedos—possibly allowing differential stratigraphic ice loss to have occurred closer to present-day (0–30 kya). However, for higher latitude NPLD exposures, IBs must have a lower albedo than expected (despite their ice-rich nature) to induce ice instability before 270 kya. This can be achieved by increasing the internal dust content of IBs or by building up/depositing dust on the surface of the IBs. We suggest the latter is more likely since radar-derived dust content for even the weakest radar reflective subsurface layers are suspected of having percentages close to that of the bulk NPLD (between 2% dust for a 1 m thick “layer”; 3% dust for bulk NPLD) (Phillips et al., 2008; Grima et al., 2009). In conclusion, albedo heavily controls the sublimation rate; however, albedo alone cannot result in positive protrusion for MBs.

MBs ablation-resistant nature is thought to be due to a higher dust content and a protective, indurated dust lag. However, whether their protruding nature could be produced from differential sublimation alone within a realistic time period (within a single prolonged excursion of solar insolation-driven ice instability) was never previously tested. We determined that the difference in dust content between a MB and the beds that sandwich it is important in causing the MBs to form as protruding layers. Higher dust content results in quicker formation of lag on the MBs. However, the lag does not need to be resistant to mechanical erosion to cause a MB to protrude.

The temporal and spatial distribution and thickness of dust cover on the cap, especially within the troughs, play an essential role in both short-term and long-term transport of water vapor within and likely out

of the polar region. We find that the key component for forming MBs that have positive protrusion is the modulation of lag production versus physical erosion of loose lag and the resulting integrated amount of time each stratum spends stable versus sublimating and producing lag. Temporal and spatial variable lag thickness between stratum types is required for MBs to protruding profiles without accounting for erosional resistance. We determined that the effectiveness of this framework on producing positive protrusion is strongly a function of dust content, which directly correlates to the amount of lag produced. Our work agrees with the recent work of Bramson et al. (2019), which suggests a few millimeters of dust on the surface is the main control on spiral trough evolution. The timeframes discussed above and those presented in Tables 6 & 7 are lower limits due to the simplified nature of our cyclical differential sublimation calculations combined with the unknown presence, distribution, and longevity of allochthonous dust-veneers. Such dust veneers pervasively cover trough walls present-day (Rodriguez et al., 2007, 2021). Trough wall sublimation can shut down entirely for an unknown number of years, even during high insolation periods if surface debris is thick enough (>cm) to insulate the ice (Bramson et al., 2019). If numerous and/or long hiatuses occurred more often than not (i.e., trough surfaces were rarely free from insulative veneers), then it's possible that trough wall ice loss (therefore, the development of MB protrusions) could have spanned multiple periods of peak instability going back 100s of thousands of years. Multiple studies suggest that the original accumulation of the upper several 100 meters of the NPLD exposed in the troughs presently could have formed 370–500 kya (Smith et al., 2016; Bramson et al., 2019). Such an age puts an upper limit on trough wall exposure and would favor troughs being insulated by veneers more often than not if trough formation (sublimation and erosion) and NPLD accumulation co-occurred. Our lower limit timeframe falls within the same timeframe as trough formation if troughs formed after NPLD deposition during the most geologically recent phase of surface erosion/sublimation 50–200 kya (Rodriguez et al., 2021). Regardless of which trough evolution hypothesis is correct, our sub-25,000-year time estimates for MB protrusion formation via sublimation and time-variable lag removal implies the importance longer-term preservation modulation of instability of entire trough wall sections via prevalent, insulative (thermally thick), veneers. Our modeled constraints stress the importance and need for more research on and spatio-temporal mapping of surface dust on the polar caps.

5.2 Spatially variable environmental conditions controlling protrusion laterally

For the purpose of this thought experiment, for simplicity (although unlikely) let us assume that the paleo-slope and orientation of a trough wall during the timeframe that ice loss last occurred was the same as present-day and that the lateral extent of the NPLD investigated at site N0 has been exposed to the surface for the same number of years. What combination of conditions would allow for the west wall to form 1.5 m of protrusion (i.e., less net ice loss) in the same amount of time as it takes the east wall to form a 5 m protrusion (i.e., more net ice loss)? As with any model, there are many possible solutions if you play around with enough variables. However, we can confidently state that the west wall would reach 1.5 m of protrusion 1,000s of years before the east wall protrusion reaches 5 m if trough wall slope and orientation was the only difference (albedo, wind speed, and lag removal rate are consistently laterally). Therefore, lateral variation in solar insolation on the trough due to trough slope and orientation cannot be the primary driver. However, slope also affects wind speed and therefore lag removal rate of loose debris. Laterally variable debris thickness is ultimately controlled by wind speed, grain size (e.g., Warner and Farmer, 2008), and degree of induration. Figure 2d in Chapter 1 shows a great example of spatial variability in veneer erosion and transport along equatorial-facing trough walls. If we assume the veneer is uniformly erosive, then the nonuniform transport of dust off the wall must be driven by variable wind speeds, possibly driven by differences in trough slope. We conclude that lateral variability in ice loss along the trough wall is best explained by lateral variability in lithic veneer/mantle/lag thickness driven by slope wind-dependent dust deposition and removal rates.

6 Conclusions

Other than erosional resistance, can you achieve the observed present-day MB topographic relief through differential sublimation alone?

- For a MB to have the chance to develop as a protrusion, the ice-rich IBs must also experience sublimation. However, bright IBs could have only experienced sublimation during the insolation peak 270 kya. Thus, IBs must have a lower albedo than expected despite their ice-rich nature (closer to that of MB albedo) to induce instability of the icy layer in more recent insolation peaks. We suggest this is most likely achieved by thermally thin dust coatings on the surface of IBs (either in the form of a thin lag via sublimation or aeolian deposition of allochthonous dust).

- 2) Temporally variable lag thickness between MB and IB is required. The key component for producing MBs with positive protrusion is the modulation of lag production versus physical erosion of the lag and the resulting integrated amount of time each stratum spends stable versus sublimating and producing lag.

What is the timespan for MB protrusion development?

- Our modeling and theoretical calculations suggest that MBs can develop into 1–5m protrusions on the order of 1000's of years. Under a range of possible surface albedos, dust content of strata, and wind speed down the equatorial-facing trough walls, we find that such topographic relief can easily be formed within a single martian precession period (e.g., high north polar insolation lasting 25,000 years driven by Mars' axial precession).
- We find that the dominant amount of ice loss from the equatorial-facing trough walls could have occurred as early as 60–125 kya. However, the speed at which topographic relief can form (1–20 kyr) leads us to suggest that thick insulative allochthonous dust veneers would be needed to force hiatuses in trough wall retreat during high polar insolation periods.

What are the driving environmental factors on spatial variability of ice loss on equatorial-facing trough slopes?

- Lateral change in solar insolation due to trough slope and orientation alone would not have been enough to cause the lateral variability in MB and IB ice loss. Nor would lateral variability in albedo (not accounting for insulative veneer). Instead, lateral variability in ice loss and, thus, present-day protrusion is best explained by the variability in slope wind speed and surface dust removal rate and their combined relationship with the differential sublimation and lag production rate between MB and IB.

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Supplementary Material

S1. Modeling diffusive loss through a lag

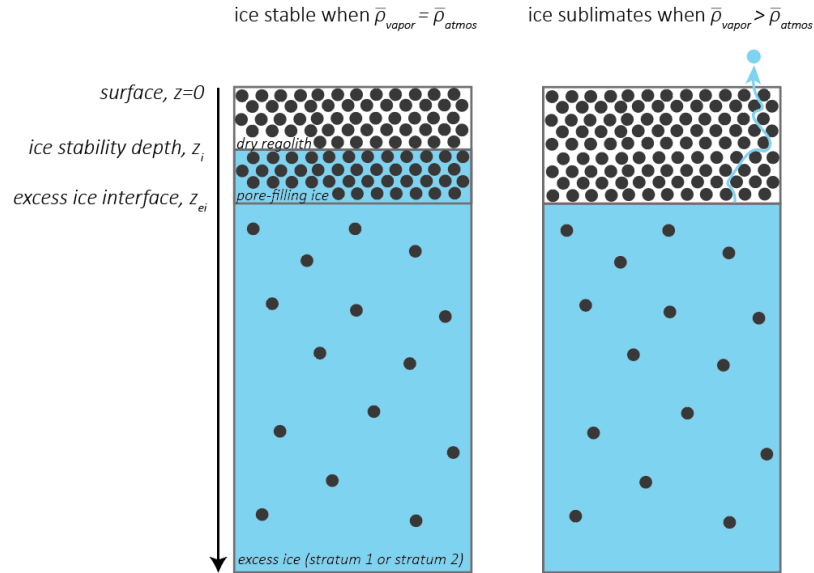


Figure S1. Graphical representation of diffusive ice loss through a regolith lag. The illustration depicts the model representation of the thermophysical layers in the case where pore-filling ice is present (exclusive to diffusive loss). See Bramson et al., 2019 and references therein for equations used for diffusive loss through a regolith lag.

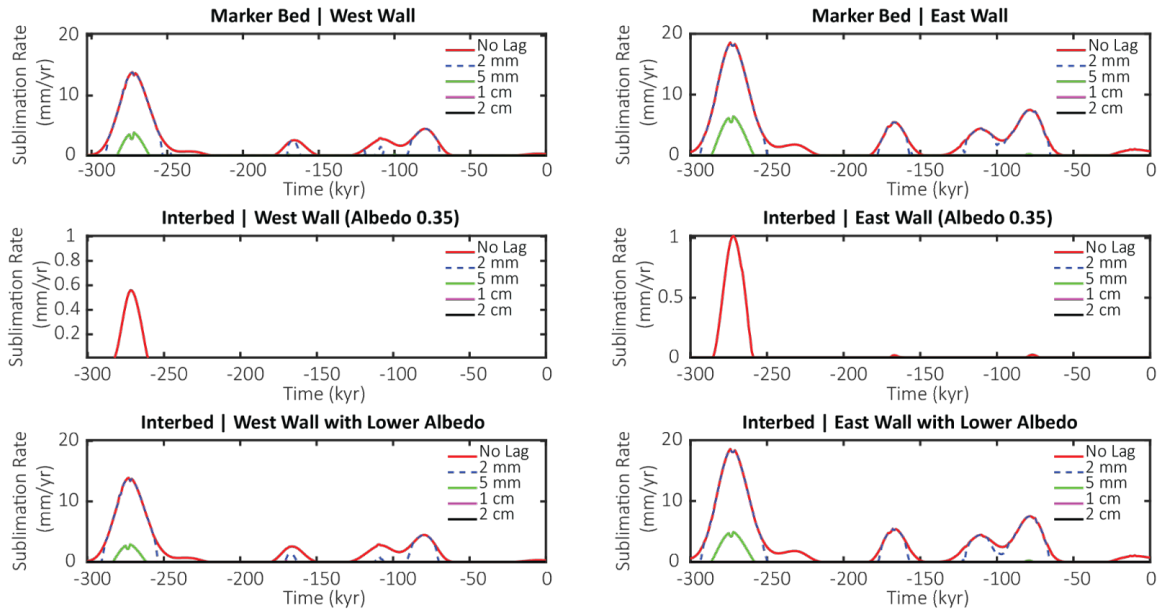


Figure S2. Instantaneous sublimation rate (mm/yr) plotted as a function of time for different lag thicknesses (no lag, 2 mm, 5 mm, 1 cm, and 2 cm) for each layer scenario (nominal water vapor and wind speed). The left and right plots show the model outputs for the west and east wall slope and orientation. The top-down order of the plots is MB, IB1, and IB2.

S2. Sublimation rates used in calculating the number of years to form MB protrusion

Table S1. Sublimation rates and lag production rates (mm/yr) used for the time estimates in Tables 6 and 7. Rates pulled from the peak modeled rates at 80 kya (Figure 7 and 8).

<i>albedo pairings</i>	MB & IB 0.15				MB 0.15 & IB 0.25		MB & IB 0.25	
<i>wind speed (m/s)</i>	0	10		20	10		10	
<i>location</i>	W	W	E	E	W	E	W	E
<i>MB sublimation rate</i>	1.8	4.5	7.5	12	4.5	7.5	0.75	1.7
<i>MB lag production rate</i>	1.3	3.2	5.3	8.5	3.2	5.3	0.55	1.2
<i>IB sublimation rate</i>	1.8	4.5	7.5	12	0.75	1.7	0.75	1.7
<i>IB lag production rate</i>	0.1	0.2	0.4	0.6	0.05	0.1	0.05	0.1

Table S2. Sublimation rates and lag production rates (mm/yr) used for the time estimates in Table 6. Rates pulled from the peak modeled rates at 110 kya (nominal runs) (Figure 7 and 8).

<i>albedo pairings</i>	MB & IB 0.15		MB 0.15 & IB 0.25		MB & IB 0.25	
<i>location</i>	W	E	W	W	E	W
<i>MB sublimation rate</i>	3	4.5	3	4.5	0.35	0.8
<i>MB lag production rate</i>	2	3.2	2	3.2	0.25	0.6
<i>IB sublimation rate</i>	3	4.5	0.35	0.8	0.35	0.8
<i>IB lag production rate</i>	0.15	0.2	0.02	0.05	0.02	0.05

Table S3. Net ice loss (in mm) perpendicular to trough surface after 1,000 years. Ice loss reported per modeled stratum for each of the assessed lag removal rates. Rates used in calculation pulled from peak modeled rates at 80 kya (Table S1)

<i>location</i>	W					E				
<i>albedo pairings</i>	MB & IB 0.15			MB 0.15 & IB 0.25	MB & IB 0.25	MB & IB 0.15			MB 0.15 & IB 0.25	MB & IB 0.25
<i>wind speed (m/s)</i>	0	10	20	10	10	0	10	20	10	10
<i>net IB ice loss (lag removal 0.05 mm/yr)</i>	600	750	715	375	375	625	760	865	570	570
<i>net MB ice loss (lag removal 0.05 mm/yr)</i>	70	70	70	70	60	70	75	70	75	70
<i>net IB ice loss (lag removal 0.1 mm/yr)</i>	900	1,505	1185	375	375	1035	1255	1510	850	850
<i>net MB ice loss (lag removal 0.1 mm/yr)</i>	120	140	140	140	105	135	140	145	145	120
<i>net IB ice loss (lag removal 0.15 mm/yr)</i>	900	1,505	1780	375	375	1035	1885	2410	850	850
<i>net MB ice loss (lag removal 0.15 mm/yr)</i>	180	200	205	195	150	195	210	215	210	190
<i>net IB ice loss (lag removal 0.5 mm/yr)</i>	900	2255	3555	375	375	1555	3760	4010	850	850
<i>net MB ice loss (lag removal 0.5 mm/yr)</i>	450	570	645	565	250	520	630	670	630	425

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Chapter 3

Sensitivity analysis of ice/dust aerosol and phase function assumptions on Hapke spectral unmixing and band depth parameters of martian water ice

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Abstract

Spectral data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM), specifically full-resolution targeted (FRT) observations, can provide quantitative information about the martian polar surface at local scales; however, it has been minimally used in such ways. This is partly due to the challenging feat of photometric and atmospheric data corrections needed, especially for icy surfaces at the poles. The scattering behavior of the dirty ice is not well-constrained. The presence of dust and ice aerosols makes it difficult to retrieve the surface's inherent albedos needed for quantitative studies of polar ice. The following work demonstrates how the limited phase angle range of CRISM “central scene” data and the assumptions made about the properties and conditions of the atmosphere (ice and dust aerosol opacity) and surface (surface phase function) going into the processing and correction of the data affect the derived surface properties of the north polar ice cap of Mars (i.e., H₂O ice band depth parameter and Hapke-modeled grain size and abundance). Through this work, we discovered the following. 1) Both dirty ice and more pure ice were best modeled as broad, moderate forward scatterers out of the four different behaviors tested. 2) The sensitivity of Hapke-modeled grain size and abundance to ice and dust aerosol opacity and phase function uncertainty is minor (well below Hapke model error). However, all possible measures should be taken to not propagate more error since Hapke modeling already inherently has large error bars (10–30%). 3) H₂O index values are affected—up to ~10% deviation—when aerosols are untreated or not treated appropriately. This should only be of concern for temporal studies. 4) Hapke-modeled grain size and abundance estimates were most sensitive to the phase angle range used for the retrieval of SSA.

1 Introduction

Although quantitative analysis of ice and dust abundance, and their grain sizes, are fundamental to understanding modern Mars polar processes, data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) full-resolution targeted (FRT) observations (18 m/pixel) (Murchie et al., 2007), well suited to this work, have been minimally used for quantitative polar cap studies. For example, the ratio of water ice, dust, and other ice impurities of specific layers of the North Polar Layered Deposits (NPLD) have been moderately constrained by ground-penetrating radar (Phillips et al., 2008; Lalich & Holt, 2017; Lalich et al., 2019). However, the direct connection between optically distinct layers and the radar data's stratification is

still a work in progress (Christian et al., 2013). As for the surface composition, grain size and water ice to dust ratios, both spatial and temporally, remain largely unconstrained by short wave infrared (SWIR; $\sim 1.0\text{--}2.5\ \mu\text{m}$) spectral data (Langevin et al., 2005; Douté et al., 2007; Calvin et al., 2009; Brown et al., 2012, 2016). Compositional information derived from hyperspectral data of individual NPLD layers would provide a useful puzzle piece in connecting the radar and visible stratigraphy to unravel the climate record. Furthermore, constraining these surface properties and their temporal/spatial variability is a cornerstone for understanding the seasonal and annual volatile mass balance and dust cycle, especially when determining what is recycled back into the polar cap versus what leaves the system (Diniega & Smith, 2019; Banfield, 2020).

However, quantifying surface dust and ice contents/grain size through the most common methods such as calculating the H_2O index (Langevin et al., 2007; Brown et al., 2010, 2012, 2016) or "spectral unmixing" through radiative transfer modeling (Langevin et al., 2005; Brown et al., 2008; Calvin et al., 2009) rely on inherent surface albedo or single-scattering albedo. Separating the apparent surface albedo (i.e., the albedo observed given a particular atmospheric state) from the inherent surface albedo is no simple feat. This is especially true over the martian poles, where dust and ice are present both in the atmosphere and on the surface, in addition to the high solar incidence angles increasing the path length of solar radiance through the atmosphere.

It has been common practice to assume polar surfaces scatter visible-near infrared radiation like Lambertian reflectors (i.e., isotropic scatterers) and to leave aerosols untreated. The treatment of aerosols requires complex and computationally expensive modeling that most scientists have limited access to such as Discrete Ordinate Radiative Transfer (DISORT) (Stamnes et al., 1988; Thomas & Stamnes, 2002) routines optimized for Mars' data, or Multi-angle Approach for Retrieval of Surface Reflectance from CRISM Observations (MARS-ReCO) (Ceamanos et al., 2013; Fernando et al., 2013, 2015; Schmidt & Fernando, 2015). The state of the atmosphere and the single-particle scattering behavior (i.e., the single-particle phase function) of the planetary surface affect the spectral albedo, spectral slope, and contrast of the surface spectra (S. Wiseman, 2009; S. M. Wiseman et al., 2016). Aerosol optical depth varies as a function of space and time, which can cause anomalous variations in surface characteristics from spectral data. These small variations can affect our interpretation of seasonal ice cap changes (e.g., frost deposition, sublimation, dust

removal) (Vincendon et al., 2007). However, it is not just aerosols that make analyses difficult. The unknown scattering properties/phase function of these dirty, icy surfaces on Mars add complexity to separating the radiance contributions from the surface and atmosphere. In the case of spectral data at the poles, ignoring aerosols in atmospheric corrections of CRISM (and other hyperspectral data)—combined with the clean/dirty ice cap's unknown phase function—may be introducing more error in our interpretations of the surface ice.

This paper aims to address the following concerns and questions: how much do assumptions about the surface and atmosphere affect analyses of remotely sensed data of polar surfaces, where both the surface and atmosphere contain ice and dust? Are the errors systematic, causing biases in the analyses? How do the assumed aerosol opacity and phase function of the surface affect retrieved polar ice/dust spectra, and how much does the uncertainty in the assumption affect the interpretation of the surface composition and grain size from the spectra? How sensitive are our spectral analyses of polar ice to our data processing techniques' rigor and assumed atmospheric and surface conditions?

Few past studies have attempted to assess the effects of aerosols or have attempted to account for their effects towards their SWIR spectral analyses (Vincendon et al., 2007; Brown et al., 2016). For example, band depth parameters, such as the H₂O ice index mapping (Langevin et al., 2005)—which uses the depth of the 1.5 μm ice absorption—is a common type of spectral analysis been used to investigate seasonal ice changes of the North Polar Residual Cap (NPRC; also referred to as NRC and NIRC) (Brown et al., 2012, 2016; Appéré et al., 2011). Brown et al. (2016) estimated the H₂O index of ice-bearing surfaces changes ~ 0.01 – 0.02 due to modest ice aerosol opacities for clear atmospheric conditions. However, this error estimate did not model the effects of dust aerosols on the H₂O index, nor were the effects of scattering behavior on the spectral analysis, which is typically assumed to be isotropic to simplify the modeling. Vincendon et al. (2007) investigated the effects of treating versus not treating dust aerosols on Observatoire pour la Minéralogie, l'Eau, les Glaces, et l'Activité (OMEGA) data over the north polar region and found that changes in seasonal ice albedo are partially controlled by variability in dust aerosol optical depth (second to the removal of surface dust by winds). Although some previous spectral studies have attempted to account for aerosols, the sensitivity of the aerosol optical depth and scattering behavior assumptions, combined, to the spectra and spectral interpretations has not been well characterized.

Why is it important to account for photometric effects due to the surface reflection characteristics and observation viewing geometry? Typically, when working with CRISM data, the radiance is converted to photometrically-corrected I/F, meaning the data are divided by the cosine of the incidence angle (also referred to as the reflectance factor). In the martian remote sensing community, it is common to assume the surface scattering behavior is isotropic (or Lambertian). For a perfect Lambertian surface, the reflectance factor does not depend on lighting and observation geometries. However, ice, snow, and frost do not behave as Lambertian reflectors but instead are highly anisotropic (Middleton & Mungall, 1952; Warren, 1982; W. Li et al., 2007; Dumont et al., 2010). For example, terrestrial snow has a strong forward scattering peak (Dumont et al., 2010). The exact anisotropic scattering behavior of the pure NPRC and dirty-ice of the NPLD are not known. Additionally, the scattering behavior of dust-covered and frost-covered ice are expected to be different from dust/frost-free ice and dirty-ice. Therefore, if the data are photometrically corrected assuming isotropic scattering, any resulting spatial differences in surface properties due to this assumption might be misinterpreted; however, the magnitude of propagated uncertainty has yet to be investigated.

The following work demonstrates how our assumption about the properties and conditions of the atmosphere and surface in our corrections of SWIR CRISM hyperspectral data influence the interpretations of surface properties of the north polar ice cap of Mars, specifically of the NPRC and NPLD. To address these concerns and questions posed in the introduction, we used the DISORT radiative transfer modeling (Stamnes et al., 1988; Thomas & Stamnes, 2002) routine optimized for the martian atmosphere (Wolff et al., 2009) to retrieve atmospherically- and photometrically-corrected single-scattering albedo (SSA) image cubes (Arvidson et al., 2006). In this paper, we describe the influences of various factors in the photometric and atmospheric DISORT correction used to calculate surface SSA from calibrated CRISM hyperspectral I/F data. Single-scattering albedo is the albedo of a material based on its scattering efficiency. It is independent of viewing geometry, which means SSA should be constant at a given wavelength regardless of the angle between the light source, surface, and detector (i.e., the phase angle). Retrieval of SSA is necessary for first order quantitative analysis of planetary surfaces, such as Hapke spectral unmixing/modeling. The novelty in the work presented herein is the employment of the multi-angle CRISM EPF (Emission Phase Function) observations in the processing pipeline, more specifically, of data acquired over the polar cap. The use of EPFs offers CRISM's full range of phase angles to SSA retrieval ($\sim 45\text{--}120^\circ$). This is different from past studies,

which only used the limited emission and phase angle ($\sim 60\text{--}85^\circ$) provided from the near-nadir, central CRISM image (also colloqually known as the “butterfly” image) to retrieve SSA (e.g., Liu et al., 2018). In this paper we refer to central scene SSA retrieval as the “simplified” SSA retrieval method or simplified approach for short. It is known and has been demonstrated that limiting the emission and phase angle range used in retrieval of SSA leads to greater uncertainty in SSA retrieval (e.g., Schmidt & Fernando, 2015). However, the propagation of this uncertainty into spectral analysis of CRISM is unknown. Therefore, in addition to assessing the affects of atmospheric and surface assumptions, this paper will also assess the effects of using the full range of viewing angles offered by the EPF versus the simplified approach of only using the central scene (Section 2.2.4 and 4.5).

We have assessed the sensitivity of SSA retrievals (and subsequent analysis/interpretation) under different aerosol and phase function parameters using DISORT, plus the sensitivity of retrievals using the DISORT-EPF approach versus the simplified approach, to answer the following:

- How do untreated aerosols affect the spectral characteristics and interpretation of the ice-rich surface through typical analyses such as H₂O Index parameters and Hapke mixture modeling?
- How does assuming isotropic versus anisotropic scattering affect spectral characteristics and interpretation of the ice-rich surface (via H₂O Index and Hapke mixture modeling)?
- How does using a DISORT-EPF SSA approach - which utilizes the full viewing geometry of CRISM EPFs - change the spectral results compared to the more common and simplified SSA retrieval methods that only use I/F (radiance at the sensor (I) divided by F, which is the incident solar irradiance divided by π) and viewing geometries from the near-nadir, central CRISM image?

2 Data set and processing

2.1 CRISM data and study site

The CRISM observation used in this paper is FRT0000BEB8, located at spiral trough site N0 (Fishbaugh et al., 2010; Becerra et al., 2016, 2017), $87^\circ\text{N } 96^\circ\text{W}$. This scene was selected for several reasons. It captures a region of the youngest, perennial ice layer known as the NPRC, the trough wall, and the trough floor. The surface coverage allows for the spectral analysis of both dirty ice and relatively bright, clean ice. Additionally, N0 has exceptional stratigraphic exposure of the upper NPLD, which has been extensively

studied and analyzed using HiRISE and CTX to identify and map layers (Fishbaugh et al., 2010; Becerra et al., 2016, 2017). The observation was acquired during optimal instrument operating temperatures (-152°C), providing high signal-to-noise (observation year 2008 or Mars year 29), and in martian northern summer ($L_s=108^{\circ}$) to avoid the seasonal CO_2 frost and dusty atmospheric conditions.

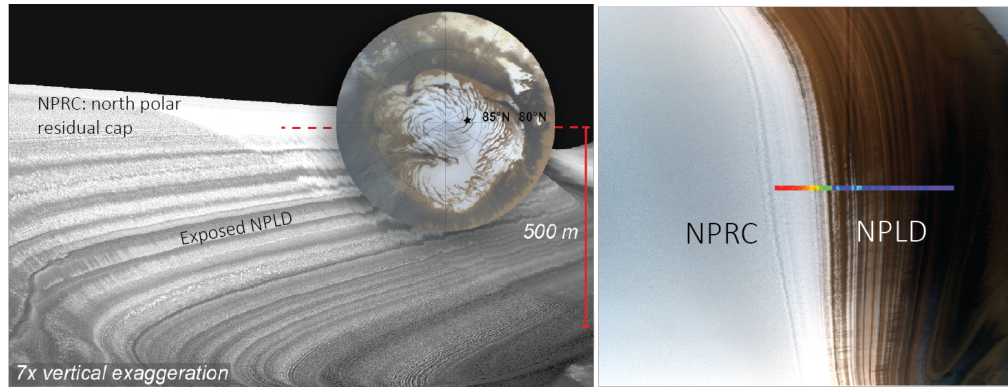


Figure 1. (left) Location of the spiral trough site N0 marked with a black star on the polar projected MARCI image. The background is CTX image overlain by the HiRISE image at N0 projected in a prospective 3D oblique view showing the exposed NPLD along the trough wall and the nearly pure ice of the NPRC. North is towards the left of the image. (right) Unprojected CRISM image at site N0 displayed in true color RGB. The rainbow strip in the middle of the image is the subset of pixels used for the spectral analysis, specifically H_2O index and Hapke mixture modeling (Sections 4.2–4.4). CTX: P02_001630_2730_XI_87N265W; HiRISE:PSP_001738_2670; CRISM: FRT0000BEB8.

2.2 Radiative transfer modeling of aerosols and surface using DISORT

The radiance received by a detector has contributions from both the surface and atmosphere. In order to separate the atmospheric aerosol and surface radiance contributions from CRISM spectra, Discrete Ordinate Radiative Transfer (DISORT) radiative transfer modeling was used (Stamnes et al., 1988; Thomas & Stamnes, 2002) and has been optimized for the martian atmosphere (Wolff et al., 2009). The DISORT model can account for the attenuation and scattering of solar radiation by gases and aerosols on its way through the atmosphere towards the surface, its interaction with the surface, and its attenuation and scattering back up through the atmosphere to the detector. DISORT-based atmospheric correction routines for CRISM and OMEGA have been developed and tested through numerous studies over the past decade for both analysis of atmospheric constituents (e.g., Smith et al., 2009, 2018) and correction for the atmosphere in order to analyze the planetary surface's spectral characteristics and properties (Arvidson et al., 2006; Cull et al., 2010; Liu et al., 2012; Shaw et al., 2013; Liu et al., 2016; Kreisch et al., 2017; Lapotre, Ehlmann, Minson, et al., 2017; Liu et al., 2018).

Before DISORT modeling is performed, atmospheric gas bands from CO₂ were removed to first order by ratioing CRISM I/F spectra to scaled atmospheric transmission spectrum derived from the Olympus Mons high and low altitude observations, also referred to as the volcano-scan method (P. C. McGuire et al., 2009). This was done using the CAT (CRISM Analysis Tool) add-on toolbox in ENVI. Gases were not explicitly modeled with DISORT due to the added computational burden. Since this work focuses on the effects of aerosols and scattering behavior on the retrieved spectra, removal of atmospheric gas contributions with the volcano-scan method to be adequate for the goals of this study. Improvements with modeling gas absorptions within DISORT compared to the volcano scan is a second effect outside the scope of this study.

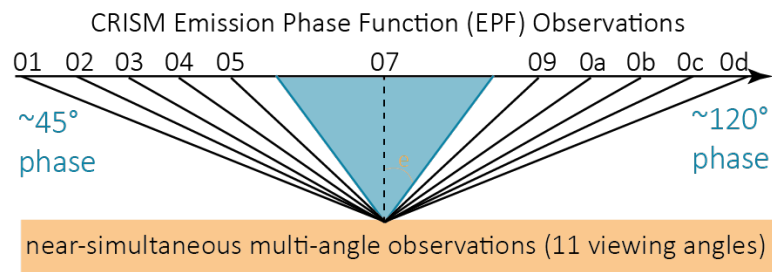


Figure 2. Schematic showing the acquisition of targeted CRISM observations and EPFs (denoted by the angled lines) (Murchie et al., 2007). Emergence and phase angle change as CRISM is flown over an area of interest on the surface (black arrow). There are 11 images acquired for one targeted observation. EPFs 01–05 are taken as the craft approaches the target, with the central image (07) being the near-nadir swath, and EPFs 09–0d are taken after the instrument has performed its full gimbal and is looking back at the target as the craft continues its orbit.

Most studies that have applied DISORT to CRISM data only used the near-nadir ($\sim 0^\circ$ emergence angle), central observation (07) (Kreisch et al., 2017; Liu et al., 2012, 2016, 2018), which only have a phase angle range from ~ 60 – 85° . The CRISM instrument’s gimbaling alters the emergence angle of each CRISM frame as it flies over the ground target. This observation mode results in a unique data set of 11 near-simultaneous multi-angle observations of the target we call Emission Phase Functions (EPFs) (Figure 2). Therefore, the EPFs give I/F values at more extreme phase and emergence angles ranging from 45° – 120° and 0 – 75° , respectively. The wide range of phase angles are exceptional for constraining and removing the scattered atmospheric light and retrieving SSA compared to only using the central CRISM observation (07) (Schmidt & Fernando, 2015). Since, at a given wavelength, the SSA should be constant at that wavelength regardless of the viewing angle, we can assess the SSA retrieved from the same location across all the EPFs,

including the central scene. By doing so, we can assess how realistic the retrieved SSA is to the true SSA as we test different opacities and phase function DISORT inputs.

The steps taken in the DISORT modeling were: 1) define an appropriate range of aerosol opacities and phase function assumptions for the CRISM observation (Table 1) (Section 2.2.1); 2) simultaneously model atmospheric and surface radiance contributions to generate forward-modeled radiance received at the sensor (I/F) for each EPF; 3) run the EPF observations through an IDL routine to retrieve surface SSA from the forward-modeled I/F values using a look-up table approach. These image cubes are then map-projected in ENVI/IDL using the CRISM Derived Data Record (DDR) information. Steps (2) and (3) are repeated for the different sets of parameters in Table 1.

Within the DISORT modeling routine, the atmosphere is parameterized, and the surface boundary is defined. The icy-dusty surface's phase function and the atmosphere's opacity are both unknown. As a result, we have four free parameters: atmospheric dust (τ_d) and water ice optical depths (τ_i), the fraction of forward-scattered light off the surface (c), and the width of the scattering lobe (b) (Figure 3). For details on the additional constraints outside this study's scope and not parameterized in the model, see Appendix A. Due to the computational expense of each DISORT-EPF SSA run, we bound the problem with seven parameter sets, which cover a range of probable opacities for the northern hemispheric summer and scattering properties one would expect in nature (Table 1). These parameter sets were chosen to assess the SSA retrieval sensitivity and the sensitivity of scientific results from spectral analysis due to unconstrained opacity and phase function.

Since aerosol optical depth and scattering behavior are both unconstrained, we delegated one parameter set as the control (Table 1, Case #3), which shares either the same phase function input (Table 1, Case #5–7) or optical depth input (Table 1, Case #2–4) as the other parameters sets (except for the isotropic-no aerosol run; Table 1, Case #1). We held τ_d and τ_i constant in order to determine which phase function best describes the ice-rich and dirty-ice surfaces of the polar cap (Table 1, Case #2–4) then assessed the SSA retrievals from the various aerosol optical depth runs while holding the phase function constant (Table 1, Case #3, 5–7).

2.2.1 Parameterizing the atmospheric model

Two variables, τ_i and τ_d , were parameterized in the atmospheric model in DISORT. The single-scattering albedo and particle phase function for martian ice and dust aerosols were used (R. Todd Clancy et al., 2003; Wolff et al., 2009) to model the extinction due to aerosols in the atmosphere. Due to the range of off-nadir to nadir geometries of the EPF observations, it is technically possible for the optical depths of aerosols to be retrieved, but only if the phase function of the surface is known or assumed, as done by Wolff et al. (2006, 2009). This type of retrieval is challenging because the atmospheric aerosols' photometric curve is often correlated with the anisotropic photometric curve of the surface materials, especially for ice and dust-rich surfaces. Others have used TES-derived historical climatology optical depths. However, the data only span from Mars year 24–26 (Earth years 1999–2003) and may not represent the atmospheric conditions during the acquisition of the CRISM scene, which did not begin acquiring data until after 2006.

Our approach assumes both the optical depths and the phase function of the surface materials are unknown. Here we used the Mars Climate Database (MCD) (Madeleine et al., 2009) to select a range of τ_d that correspond to the expected minimum and maximum dust column optical depth for martian northern summer between $L_s = 80$ – 160° at the location of FRT0000BEB8 (Table 1, Case #5 & #6). The MCD interpolates optical depth retrieval estimates from TES and THEMIS nadir observations and Mars Climate Sounder limb observations, covering more recent years than the historical TES climatology data. This range was chosen to assess how the uncertainty in knowing τ_d , due to locally unpredictable atmospheric dust variability, would affect the spectral analysis. In addition to the min-max range, we used the approximated τ_d for the specified Mars year, L_s , local solar time, latitude, and longitude of FRT0000BEB8 (Table 1, Case #3), which happens to fall between the minimum and maximum τ_d estimates. We set τ_i to 0.0 (tagged at $12.1 \mu\text{m}$) for all runs except for one, which we set to 0.05 (Table 1, Case #7) because there is expected to be a minimal amount of ice aerosols during mid-northern hemispheric summer. Additionally, the sensitivity of τ_i on H_2O indices has previously been done by Brown et al. (2016), which is discussed in Section 5.

2.2.2 Parameterizing the surface boundary

The surface boundary within DISORT was defined by the Hapke bidirectional reflectance function (Hapke, 1993, eqn. 8.89) incorporating the double Henyey-Greenstein (HG2) single-particle phase function (A. F. McGuire & Hapke, 1995), which was used to retrieve surface SSA using the c and b scattering behavior/phase function parameters. These parameters describe the angular width of the light being scattered (b) and the fraction of light that is being scattered in the forward direction away from the light source versus backward. The equations used are explained in-depth below. Surface-scattering is defined according to the Hapke bidirectional reflectance equation (Hapke, 1981, Eqn. 16'; Hapke, 1993, Eqn. 8.89). The bidirectional reflectance is then converted to a radiance coefficient (Hapke, 1981, Eqn. 37) (the bidirectional reflectance relative to a diffuse standard illuminated and viewed normally) by multiplying the values by $\pi/\cos(\text{incidence angle})$ (Hapke, 1993, Eqn. 10.4). This reflectance conversion is the equivalent to CRISM I/F. And then the radiance coefficient is multiplied by S , the shadowing function, defined by a macroscopic roughness parameter, $\bar{\theta}$, which describes the average topographic slope angle of surface roughness at the sub-resolution scale relative to the direction of illumination (Hapke 1993, Eqn. 12.54) for when $e \leq i$. The final form of the equation used to define the surface can be written as follows:

$$r_c(i, e, g) = \frac{\omega_{ave}}{4} \frac{1}{\mu_0 + \mu} \{ [1 + B(g)]P(g) + H(\mu_0, \omega_{ave})H(\mu, \omega_{ave}) - 1 \} S(i, e, g)$$

In the above equation, i, e , and g are the local MOLA derived incidence, emergence, and phase angles, respectively, retrieved from the CRISM observation's associated DDR file. μ_0 is the cosine of the incidence angle and, μ is the cosine of the emergence angle, $B(g)$ is the opposition effect (describes the surge in brightness close to zero phase angle), $P(g)$ is the single-particle phase function, and $H(\mu_0) H(\mu)$ is the multiple scattering H-function. Equations for the functions not described in the main text below can be found in Appendix A.

Here, we are assuming an anisotropic surface; therefore, the phase function is modeled as a double Henyey-Greenstein function (HG2), which is dependent on the phase angle and two scattering parameters:

$$P(g) = c \frac{(1 - b^2)}{(1 + 2b \cos g + b^2)^{3/2}} + (1 - c) \frac{(1 - b^2)}{(1 - 2b \cos g + b^2)^{3/2}}$$

, where c is a weighting factor that describes the fraction of forward-scattering or magnitude of the backscattering lobe relative to the forward scatter lobe (1 = all forward scatter, 0 = all backscatter) and b is the asymmetry factor, which describes the degree to which the phase function is anisotropic or the angular width of the scattering lobes and is constrained between 0 and 1 (1 = isotropic scatter).

We chose four different combinations of c and b to test a range of different scattering behaviors, including isotropic, wide backscatter, modestly forward scatter (that of martian regolith), and narrow forward scatter (most akin to snow). Figure 3 is a plot of c versus b that was chosen in this study relative to those derived for terrestrial snow, frost, and the surfaces of icy moons Europa and Rhea (Middleton & Mungall, 1952; Domingue et al., 1997). We assumed the same scattering behavior of the average martian regolith as Johnson et al. (2006) for the control run (Figure 3, black star).

Table 1. Atmospheric opacity and surface phase function cases

Input Parameter Cases	τ_a (@ 9.3 μm)	τ_i (@ 12.1 μm)	b (scattering lobe asymmetry)	c (fraction of forward scattering)
#1 Isotropic, aerosols not treated	0.001	0.0	0.0	0.5
#2 Wide backscatter	0.124	0.0	0.1	0
#3 Broad, modestly forward scattering (control)	0.124	0.0	0.25	0.6
#4 Narrow forward scattering	0.124	0.0	0.5	0.8
#5 Minimum summer dust estimate*	0.100	0.0	0.25	0.6
#6 Maximum summer dust estimate*	0.130	0.0	0.25	0.6
#7 Ice aerosols	0.124	0.05	0.25	0.6

* We used Mars Climate Database (MCD) climate scenario MY 29 to retrieve estimates for minimum and maximum likely summer τ_a (between $L_s = 80\text{--}160^\circ$).

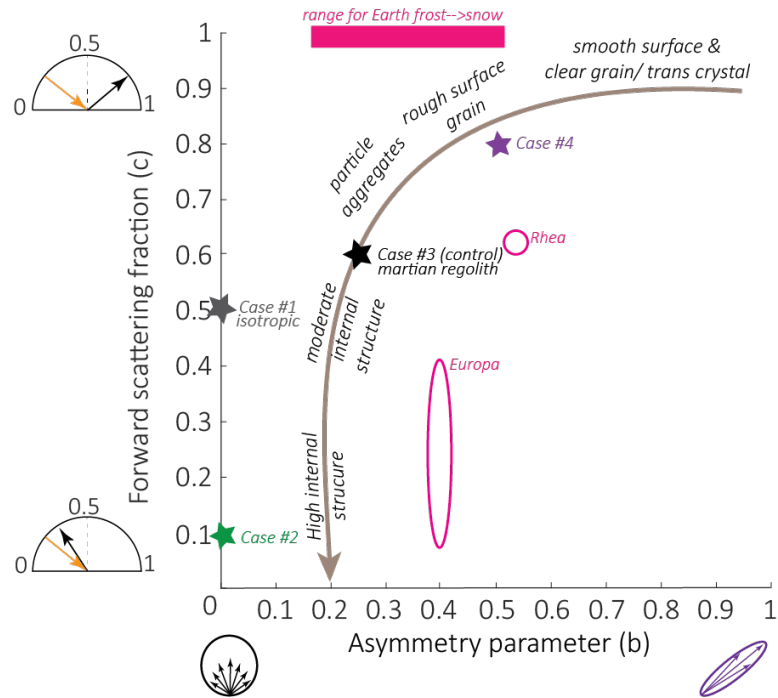


Figure 3. Adopted schematic plot from McGuire and Hapke (1995) showing the general correlation between physical lithic grain properties and their scattering behavior (brown arrow) defined by parameters c and b in the 2-lobe Henyey-Greenstein single-particle phase function. The colored stars represent the different b and c values used for the DISORT cases (Table 1, Case #1– 4) In pink, for comparison, are the plotted scattering behavior of terrestrial frost, snow and the icy surface of the moons Europa and Rhea (Domingue et al., 1997). Our control case (black star) uses the scattering behavior of average martian soil determined by Johnson et al. (2006).

Table 2. FRT0000BEB8 Observation Information

L_s	108.27
Mars Year (MY)	29
Latitude	87.047 N
Longitude	96.17 E
Local Solar Time (hour angle degrees)	11.75
UTC	Aug 05, 2008

2.2.3 Assessing the surface single-scattering albedo (SSA) retrievals

At a given wavelength, SSA is independent of viewing geometry. In other words, SSA should be constant with phase angle. The input parameter case that best approximates a constant SSA as a function of phase angle would best describe the state of the atmosphere and surface for the CRISM observation. We define a criterion for the ideal SSA retrieval in the equation depicted in Figure 4. The ΔSSA metric is calculated from the SSA-phase retrieval curve at each wavelength (Figure 4 subset plot, Figure S2). An absolute perfect retrieval of the surface's true SSA would have a ΔSSA of 0 at every wavelength. Therefore, the best retrieval case is deemed one that has a mean ΔSSA and mean standard deviation closest to zero. Mean and standard deviation ΔSSA values were recorded (Figure 5). The average spectra for each DISORT case from each EPF can be found in the supplementary material (Figure S3 for NRPC ROI and Figure S4 for NPLD ROI).

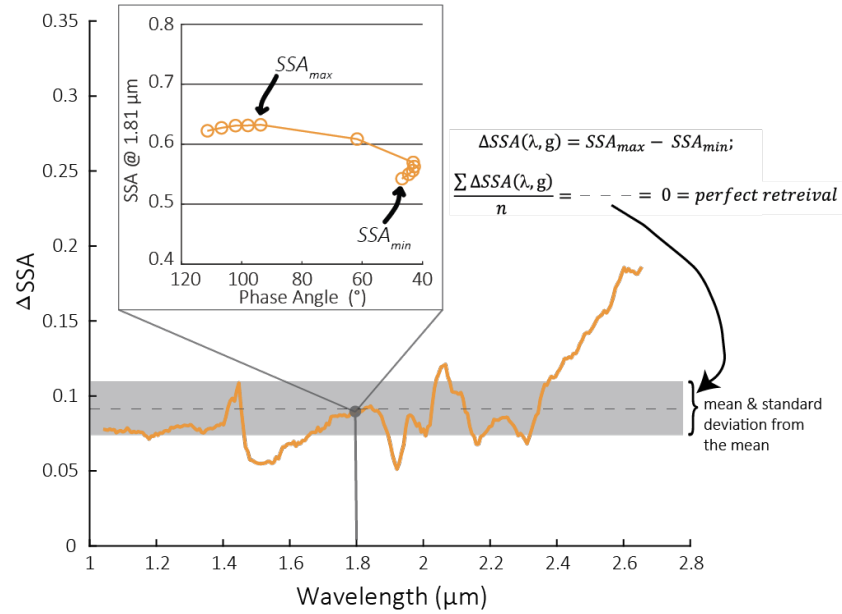


Figure 4. An example of the retrieved ΔSSA as a function of wavelength, which is calculated from retrieved SSA as a function of phase angle (see subset plot). In the criterion equation, n equals the total number of wavelengths, λ , and g equals phase angle. The subset plot (SSA-phase retrieval curve) shows an example of the retrieved SSA at $1.81 \mu m$ as a function of phase angle from Case #5. Each open circle data point in the subset is the average SSA for a single EPF (extracted from an ROI that overlaps the same spot on the surface across each EPF). A perfect retrieval would be a horizontal line for SSA as a function of phase angle and a horizontal line at 0 ΔSSA for SSA as a function of wavelength.

2.2.4 DISORT-EPF SSA retrieval method versus simplified SSA retrieval method

In addition to the seven different DISORT-EPF SSA retrievals, we performed a more simplified I/F to SSA conversion that does not use DISORT or the EPFs in its photometric correction and conversion process. A main difference between the two SSA retrieval methods lies in the range of phase angles used to create the look-up table to convert between I/F and SSA. The simplified approach only uses the central scene (observation 07—the “butterfly” image when map-projected) viewing geometry, which is limited to phase angles $\sim 60\text{--}85^\circ$, while the DISORT-EPF SSA method uses the full phase angle range $\sim 45\text{--}120^\circ$ offered by the EPFs. This approach also typically assumes isotropic scattering and that there are no aerosols to simplify the computation (Mustard and Pieters, 1987). The SSA from the simplified approach is not as accurate as methods that use the full range of phase angles offered by the EPFs (Schmidt & Fernando, 2015). In Section 4.5, we compare the spectral analysis results from the DISORT-EPF SSA Case #1 (isotropic-no aerosols) to the results from the simplified approach to assess how not using a more rigorous retrieval routine that uses EPF observations like DISORT affects the data and interpretations of the icy martian surface.

3 Spectral analysis techniques

3.1 Spectral unmixing of SSA using Hapke theory

Quantitative compositional analysis of remotely sensed planetary spectral data is most often met with the challenges of not being able to ground truth the results. However, radiative transfer modeling allows planetary scientists to obtain the most quantitative compositional information possible beyond simple identification of the presence and absence of minerals and other surface components when in-situ data are not available (Mustard & Glotch, 2019).

The DISORT-EPF SSA retrieval process produced denoised SSA image cubes for each EPF observation (Kreisch et al., 2017). The SSA image cube (observation 07—the central scene) was used to solve for abundance and grain sizes using an intimate ternary mixture model of water ice, dust, and palagonite using their optical constants (Warren & Brandt, 2008, Wolff et al., 2006, 2009). We used the following Hapke equation for mixture modeling of relative component abundance and effective grain sizes,

$$\omega_{ave} = \sum_i \frac{M_i Q_{si}}{\rho_i d_i} / \sum_i \frac{M_i Q_{se}}{\rho_i d_i}$$

, where ω_{ave} is the average SSA, M_i is the mass fraction of component i , ρ_i is the solid particle density, d_i is the average effective particle diameter, and Q_{si} is the scattering efficiency of the i th component. The summation is carried out for all components in the mixture. Here it is assumed that particles on the surface are much larger than the wavelengths, therefore the extinction efficiency, Q_{se} , is 1 (Hapke, 1981). Q_{si} is calculated using the optical constants of the i th components. The above equation is solved for the mass fraction (or relative abundance in the case of orbital data) and the effective grain size of each i th component by forward modeling. To solve the nonlinear Hapke equations, numerous mass fraction and grain size combinations are tested until the best spectral fit is achieved using a Newton's theory and Levenberg-Marquard least square nonnegative algorithm for optimization (using the *lsqnonlin* function in Matlab) (S. Li & Li, 2011; S. Li & Milliken, 2015; Robertson et al., 2016). For the full set of equations used and the parameterization of the unmixing, see Appendix B.

3.1.1 Hapke modeling caveats to keep in mind

Like most Hapke results from planetary data that lack in-situ data for validation, we acknowledge that abundance and grain size estimates presented in this work are first-order estimates and should not be taken as surface property estimates outside the context of this paper's goals outlined in Section 1. Hapke mixture modeling has been used successfully in the laboratory setting to quantify samples' mineral abundances down to 5% accuracy of their mass by weight percent (S. Li & Li, 2011; S. Li & Milliken, 2015). Such impressive accuracy using laboratory data cannot be expected from orbital data. The uncertainty in modeled abundance estimates from remotely sensed SWIR data range from <10% up to 20–30% for certain mixtures as a consequence of instrumental and atmospheric noise, imprecise exact viewing geometry, and unknowns of textural and optical properties of surface materials (Lapotre, Ehlmann, & Minson, 2017; Lapotre, Ehlmann, Minson, et al., 2017; Harris & Grindrod, 2018). However, assessing the validity of grain estimates is outside this study's scope. This study is focused on investigating the relative change in the modeled grain size (and abundance) due to model input uncertainty in the data correction process. Hapke model results from orbital data should always be treated with caution regarding relative abundance and effective grain size (even when starting with optical constants of the spectral endmembers). Despite the uncertainty in applying the Hapke model to orbital data, the nonlinear unmixing techniques such as Hapke

are still the most quantitative method for assessing compositional variability in hyperspectral data and interpreting the surface geology (Mustard & Glotch, 2019).

The abundances and grain sizes reported in this paper should not be used as new constraints on the surface's compositional and grain variability. This is a comparative study; we did not perform an in-depth analysis and assessment on the validity of the Hapke mixture model assumptions (e.g., intimate mixture vs. areal mixture or coating) used or the resulting spectral fits and component mixture estimates. A recent reflectance study about frost on ice and dust intimate mixtures show that frost coatings impact the visible spectral slope rapidly while its effect on the water ice absorptions in the SWIR is less (Yoldi et al., 2020). Therefore, the visible slope can be compared to the H₂O index to identify whether H₂O frost—even a few tens of microns thick—is coating the surface of grains. Future studies should be conducted to investigate the assumptions made about the various properties of the polar cap surface particulates and whether an intimate mixture alone is a suitable assumption for the ice caps. Areal mixtures and coatings—whether it is a dust-coated ice grain or frost-coated dust grain—are properties that future studies should investigate with laboratory and CRISM data.

3.2 H₂O index band depth parameter

The reflectance spectrum of H₂O ice has three main absorption features between 1 and 4 μm at 1.5 μm , 2.0 μm , and 3.0 μm . The H₂O index is essentially the continuum removed band depth at 1.5 μm and is sensitive to grain size and dark impurity content, making the index useful for qualitative assessments of ice purity. The 1.5 μm absorption is preferred over the more prominent 2.0 μm or the 3.0 μm as the H₂O index because the 2.0 μm and 3.0 μm absorptions become saturated quickly at grain sizes larger than a few hundred μm and a few μm , respectively (Grundy & Schmitt, 1998).

The H₂O index is calculated from reflectance data. Therefore, the denoised SSA image cube produced by the DISORT-EPF SSA pipeline was converted to reflectance using the Hapke radiance coefficient equation (Hapke, 1993). The same assumptions for scattering behavior prescribed to the phase function in the DISORT routine for a given run was also used to convert from SSA back to reflectance. The reflectance data were then used to calculate H₂O index maps of each iteration of DISORT using the following index modified from Langevin et al. (2007) by Brown et al. (2010):

$$H2Oindex = 1 - \frac{R(1.500)}{R(1.394)^{0.7} * R(1.750)^{0.3}}$$

4 Results

The result sections below will cover the following: Section 4.1, how and which set of parameters (Table 1) produced the best SSA retrievals for the NPRC and NPLD (i.e., trough wall) and which were the worst. Section 4.2 investigates how the spectral characteristics (spectral contrast and spectral albedo) of the retrieved SSA spectra change as a function of the assumed atmospheric and surface properties. Section 4.3 and 4.4 put the spectral changes, discussion in Section 4.2, into context by comparing the results from Hapke mixture models and band depth parameter analysis (i.e., H₂O index) for each DISORT case to our "control" case. In other words, how sensitive are these types of polar spectral analyses to the assumptions going into data processing? Lastly, Section 4.5 compares how using EPFs versus only using the central scene to retrieve SSA affects the H₂O index and Hapke mixture model results.

4.1 Determining the best retrieval of SSA

First, we set out to determine the set of atmospheric and phase function parameters that best represents the aerosol optical depth and the scattering behavior of the ice cap at the location and time the data were acquired, as described in Section 2.2.1, given our parameter space. Mean Δ SSA and standard deviation from the mean were used to determine which case derived SSA closest to true SSA. Lower mean Δ SSA and standard deviation imply retrieval is closer to true SSA. These values were extracted from ROIs that covered the same location on the surface across all the EPF observations (Figure S1) for each DISORT case and are presented in Figure 5. A more detailed discussion of the SSA retrievals from each case can be found in Supplementary Material S1.

Out of the seven different DISORT runs performed on FRT0000BEB8, we found that a broad-moderately forward scattering, lower τ_d , and $\tau_i = 0$ provided the best SSA retrievals for both the NPRC and NPLD (Case #5) (Figure 5). In other words, using the dust opacity estimates from the MCD (or slightly less than) for the date, time, and location of the observation, no ice aerosols, and using the average b and c values of martian regolith from Johnson et al., 2006 gave the best SSA spectral retrievals.

Our study showed that narrow forward scattering (Case #4), isotropic scattering with no aerosols treated (Case #1), and treatment of modest ice aerosol opacity (Case #7) yielded the worst retrievals (Section S1, Figure 5, Figure S2). They either have overall the highest mean Δ SSA and or greatest Δ SSA range and standard deviation. These three runs also resulted in failed SSA retrievals (i.e., some SSA values retrieved were >1.0 or <0.0 for any given phase angle) (Figure S3 and S4). Although three of the seven parameter sets tested resulted in failed SSA retrievals, we choose to include all seven in the following assessment to thoroughly discuss the full range of effects from the unconstrained surface/atmospheric parameter inputs.

Although we found a case that provides the best estimate, it is not the true single-scattering albedo of the surface. The actual single-scattering albedo of the surface cannot be accurately determined unless we have ground-truthed data from a lander or rover (e.g., using the various cameras and lighting conditions of the MER Spirit rover to get at the scattering properties of the different types of regolith (Johnson et al., 2006)). Alternatively, one could use more statistical and technical methods such as the Multi-angle Approach for Retrieval of Surface Reflectance from CRISM Observations (MARS-ReCO) (Ceamanos et al., 2013; Fernando et al., 2013, 2015; Schmidt & Fernando, 2015). MARS-ReCO uses a Bayesian inversion framework based on Hapke's model using the CRISM EPFs to retrieve estimates for the surface photometric properties, which have been cross-validated with MER rover data (Ceamanos et al., 2013; Fernando et al., 2013).

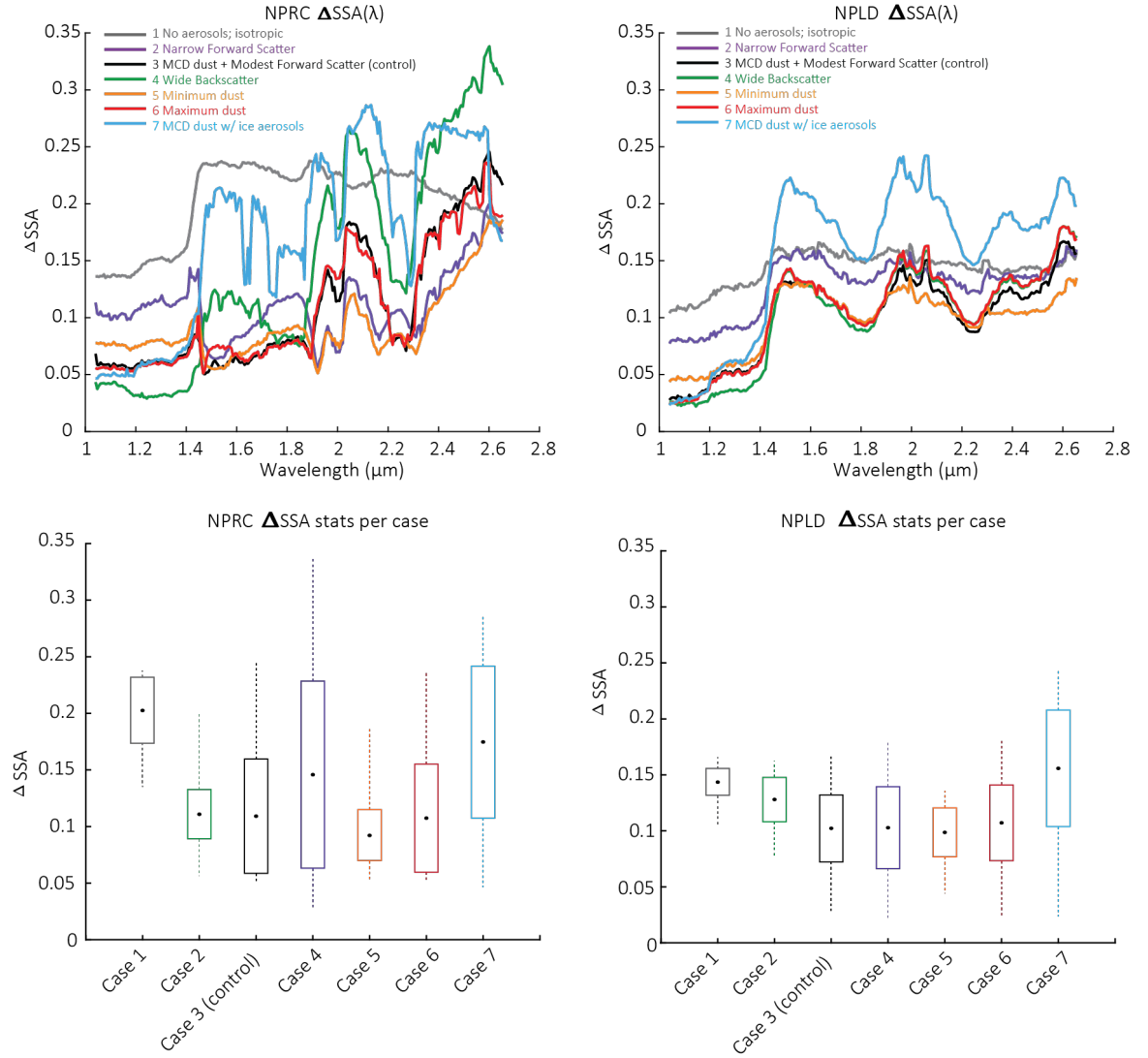


Figure 5. The top two plots show the retrieved ΔSSA as a function of wavelength for each case extracted from the ROIs in Figure S1. Refer to Figure 4 for how these plots are produced. The bottom two plots summarize the data in the top two plots by plotting the mean, standard deviation from the mean, and min-max ΔSSA values for each case. The best retrievals are cases that have smaller deviations in ΔSSA and mean ΔSSA closer to 0. Case #5 has, combined, the lowest mean and smallest standard deviation for both the NPRC and the NPLD ROIs.

4.2 Impact of aerosols and scattering behavior assumptions on spectral features

The results in Sections 4.2.1 and 4.2.2 can be summarized as follows.

The effect of unconstrained aerosol opacity on retrieval of SSA spectra:

- Spectral contrast of the pure ice and dirty ice spectra is lower when aerosols are not treated.
- Surface ice absorption band depth increases when water ice aerosols are assumed present.
- Treatment of dust aerosols also increases the water ice absorption band depths but to a lesser degree.
- There is little spectral albedo change outside of the major absorptions.
- The spectra are redder (shallower spectral slope) when dust aerosols are left untreated.
- The effect of unconstrained phase function parameters on retrieval of SSA spectra:
- Changing the assumed phase function parameters, b and c , either increases or decreases the spectral albedo.
- Spectral contrast is not affected because we assume the phase function parameters, b and c , are constant as a function of wavelength (in nature, this assumption is incorrect (Pilorget et al., 2016)).

4.2.1 Spectral effects due to aerosol opacity assumptions

When aerosols' effects are not included, the spectral contrast (depth of the absorption bands) is lower (Figure 6: e.g., contrast the spectrum of no aerosols vs. MCD $\tau_d @ 9\mu\text{m}$). When ice aerosols were treated, the ice absorption bands from the NPRC spectra become noticeably deeper. The overcorrection of ice aerosols ($0.05 \tau_i @ 12.1 \mu\text{m}$) resulted in the $1.5 \mu\text{m}$ absorption bands increasing by 4% relative to our control run. Similarly, higher τ_d results in increased absorption strength, but to a lesser degree ($<2\%$) than τ_i . This effect is much more evident in the ice-rich spectra from the NPRC. The albedo outside of the absorptions does not change substantially. Overall, ice aerosols, whether corrected or not, affect the surface spectral albedo and absorption depth more so than the dust aerosols.

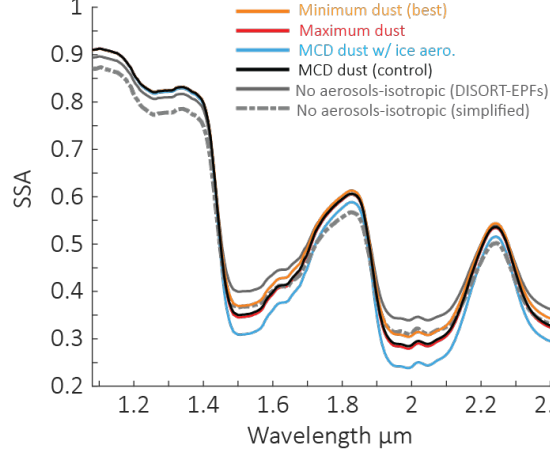
4.2.2 Spectral effects due to phase function assumptions

To assess changes in the retrieved spectral characteristics induced by the assumed single particle phase function, we kept the τ constant for three runs but changed the phase function parameters b and c . These parameters describe the angular width of the light being scattered (b) and the fraction of light scattered in the forward direction away from the light source versus backward (c). We chose three different sets of possible phase function parameters to cover a broad scope of surface types described in Section 2.2.2. Our results show that the assumed scattering behavior greatly affects the spectral albedo but does not affect the spectral contrast or spectral slope (Figure 6). This effect was anticipated since the b and c are assumed to be independent of wavelength. The DISORT routine is limited to this assumption, which decreases the complexity of teasing apart aerosol contributions versus surface-scattering. However, in nature, a material's scattering behavior varies with wavelength (Pilorget et al., 2016; Shaw et al., 2013). Constraining b and c as a function of wavelength for icy-dusty surfaces was outside this study's scope but could be a worthwhile investigation for future studies.

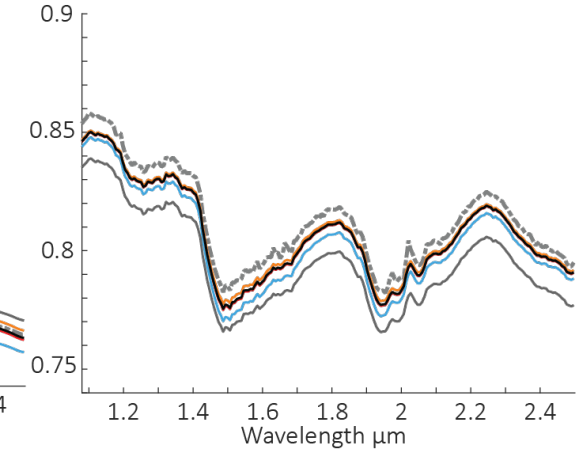
Our results show that assuming a narrow forward scattering behavior (Table 1, Case #4) increases the overall spectral albedo while a more broad and backscattering behavior (Table 1, Case #2) lowers the spectral albedo. The overall difference in spectral albedo between the backscattering and narrow forward scattering assumptions was up to 2.5% for the NPRC spectra and ~1% for the NPLD spectra. The spectral albedo for the broad modestly forward scattering (control) run lies between these two extreme scattering behaviors. The fact that the spectral albedo of the retrieved spectra is higher or lower relative to the unaffected spectral contrast is important because the spectral unmixing will attempt to match the absorption strengths and albedo by varying the estimated grain sizes and abundances of ice, palagonite, and martian dust. The implications this has on the interpretation of the surface ice, and ice impurities are discussed in Section 4.3 & 5.1.

Spectral effects due to aerosol assumption

(a) NPRC

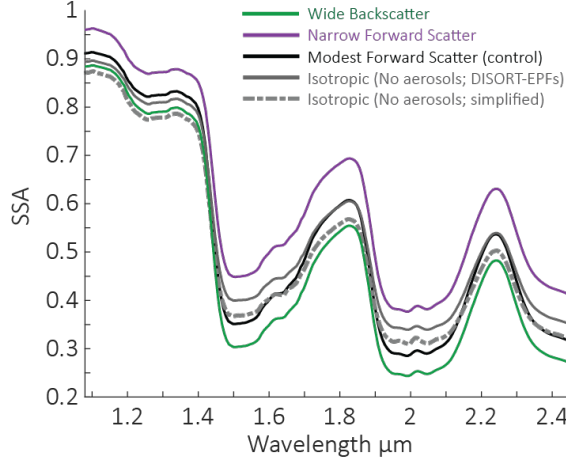


(b) NPLD



Spectral effects due to phase function assumptions

(c) NPRC



(d) NPLD

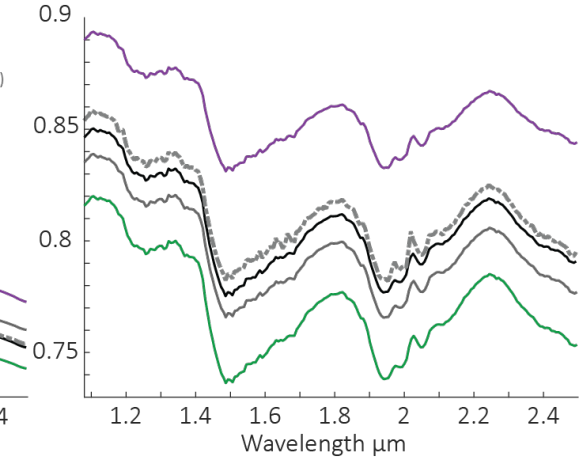


Figure 6. Spectral plots showing the average ROI single-scattering albedo spectra of the NPRC (a and c) and the trough wall from FRT0000BEB8. ROIs were 100,868 pixels and 52,207 pixels respectively, covering as much of the NPRC and trough as possible in the central CRISM scene. Plot (a) and (b) show the average spectra from the central CRISM observation for each of the different aerosol assumptions (from the same NPRC and trough wall ROI) in order to show the different spectral effects the aerosol assumption has (see colored legend). Plot (c) and (d) show the average spectra from the central CRISM observation for each of the different phase function assumptions in order to show the different spectral effects the assumed phase function parameters have on the retrieved SSA spectra (see colored legend). MCD dust = dust opacity estimated from the Mars Climate Database for the time and location of the CRISM observation.

4.3 Sensitivity of Hapke Results

The sensitivity of aerosol optical depth and phase function parameters on mixture modeling results can be summarized as the following:

- Aerosol optical depth has a negligible effect on ice abundance estimates (Figure 7a).
- The change in spectral contrast induced by the different aerosol correction cases affected the estimated grain size more than it did the abundance (Figure 7c).
- The change in spectral albedo induced by varying the assumed phase function of the icy surface has the greatest effect on estimated ice abundance (Figure 7b).
- The increase/decrease in estimated ice abundance due phase function's effect on albedo also affect the estimated grain size. In other words, driven by the albedo, a balance is met between abundance and grain size to produce the best spectral fit (Figure 7d).
- Overall, Hapke's sensitivity to the assumed phase function parameters and opacity inputs through DISORT is within the inherent model uncertainty of Hapke.

4.3.1 Sensitivity of nonlinear spectral modeling to optical depth uncertainty

Figure 7a and 7c plot the comparative model ice abundance and grain size between the different opacity cases. We found that uncertainty in knowing aerosol τ resulted in negligible variation in surface ice abundance estimates (Figure 7a). When the dust and ice τ are varied (Table 1, Case #5, 6, and 7), 95% of ice abundance estimates deviate $<3\%$ from the control case (Case #3), which is well below the 10% Hapke spectral modeling detection limit (Lapotre, Ehlmann, & Minson, 2017; Lapotre, Ehlmann, Minson, et al., 2017). Getting deviations in abundance estimates under the model's theoretical detection limit suggests that the differences observed in spectral contrast induced by the varied τ_i and τ_d (Figure 6a, 6b) are insignificant.

As shown in Figure 6, the expected northern summer dust opacity range (from the MCD) affects the spectral contrast at most up to 2%. This difference in band depth, induced by the correction for aerosols, affected the estimated grain size more than it did abundance. The $\sim 4\%$ increase in the ice absorption band depths was enough to make the Hapke model estimate larger ice grain sizes relative to the control run due to the deeper ice absorption bands from overcorrection of the ice aerosols, whose effect can be seen in Figure 7c. For Case #7 ($\tau_i = 0.05$), 95% of the grain size estimates deviated $<9.7\%$ from the control case's estimates.

For our maximum summer dust case (Case #6), 95% of the data deviated $<3.2\%$ from the control case's grain size estimates. Li and Milliken found that when inverse Hapke modeling was applied to sieved HED samples, the modeled grain sizes were larger ($>2x$) than that of the average reported sieve range (S. Li & Milliken, 2015). Therefore, we believe the sensitivity of Hapke model estimates for grain size to opacity is well within the Hapke error.

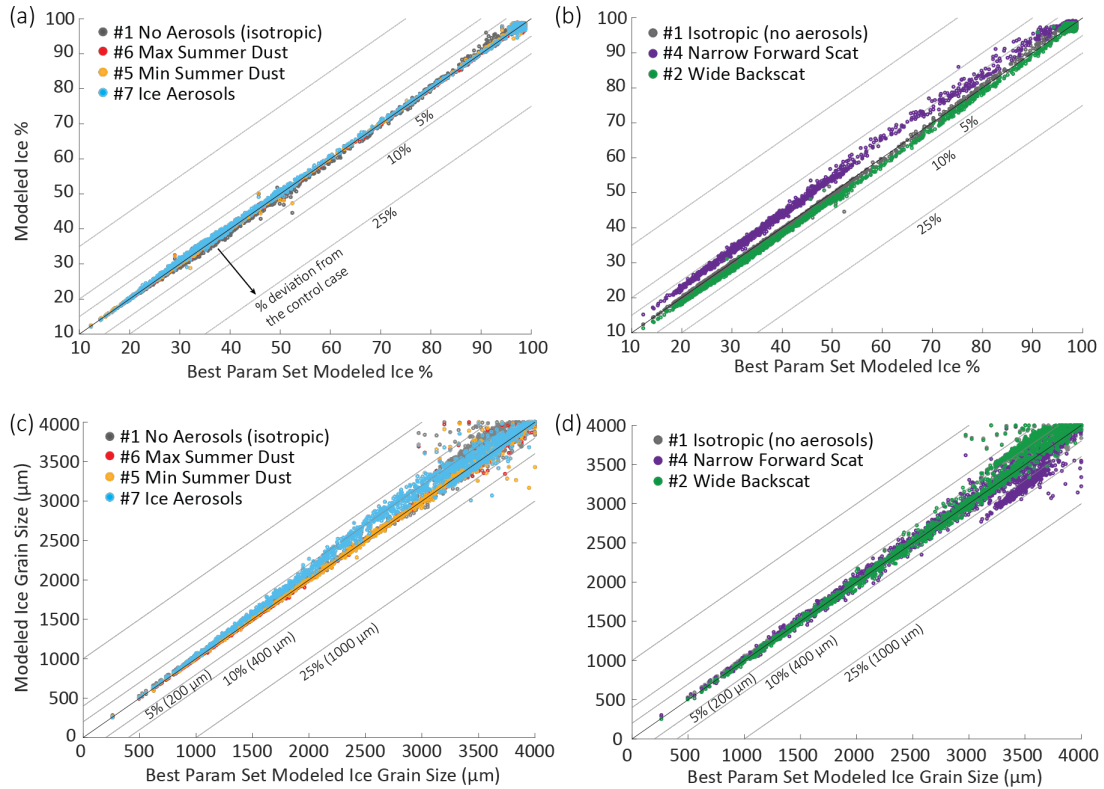


Figure 7. The plots above compare the Hapke unmixing results between the different parameter cases. The scatter plots compare the model results from the control case—Table 1, Case #3—(x-axis) to all the other cases (y-axis). Results that are close to the control case plot tightly to the one-to-one line (solid black line). The grey lines labeled 5%, 10%, and 25% mark the percent deviation from the control case. The left plots (a & c) compare the cases where τ was varied while the scattering behavior was held constant. The right plots (b & d) compare the cases where scattering behavior was varied while τ was held constant.

4.3.2 Sensitivity of nonlinear spectral unmixing to phase function parameter uncertainty

Figure 7b and 7d plot the comparative model ice abundance and grain size between the different phase function parameter cases. The phase function controls the overall albedo of the retrieved SSA spectra (Figure 6c, 6d). That change in albedo has the greatest effect on ice abundance, specifically for the results

from the narrow forward scattering correction, which produced the brightest spectra. This result implies that the model adjusts the ice abundance to compensate for the change in spectral albedo due to the different scattering assumption cases. The modeled ice abundances from the narrow forward scattering case (Table 1, Case #4) are 3% higher on average, with 95% of all the cases estimates deviating $<6.5\%$ from the modest forward scattering control case (the one-to-one line). Due to the higher albedo of the spectra from Case #4, the estimated ice abundances from the Hapke model increase relative to the other cases whose spectral albedo are lower, which favors slightly lower ice abundances and higher dust abundance. For example, the modeled ice abundances from the modest backscattering case (Table 1, Case #2) are $<4.5\%$ lower than control case estimates. The modeled ice abundances from the isotropic case (Table 1, Case #1) had a mean deviation close to zero, with 95% of its data deviating $< \pm 2\%$ from the control estimated abundances despite being one of the worst SSA retrievals (Section 4.1 & Supplementary Material S1). Overall, the variation in modeled ice abundance estimates across all scattering cases is well within the 10–30% Hapke spectral modeling detection limit (Lapotre, Ehlmann, & Minson, 2017; Lapotre, Ehlmann, Minson, et al., 2017).

The difference in ice abundance was not the only effect observed when scattering assumptions were varied. The change in albedo between the scattering cases also affected the grain size estimates. To find the best spectral fit that would match the high albedo spectra as well as the spectral contrast from the narrow forward scattering case (Table 1, Case #4), the model favored higher estimates in ice abundance, which was then compensated by lower ice grain size estimates. A similar but inverse effect occurred for the lower albedo retrievals from the broad backscatter case (Table 1, Case #2). In general, the deviation in the scattering cases' ice grain size estimates from the control case was similar to the deviation caused by the inclusion of τ_i . Following the logic in Section 4.3.1, we believe the sensitivity of Hapke model estimates for grain size to scattering behavior is also within the Hapke error.

4.4 Sensitivity of the H₂O index

The sensitivity of aerosol optical depth and phase function values on band depth parameters (i.e., 1.5 μm H₂O Index) from retrieved reflectance data can be summarized as follows:

- Uncertainty in aerosol optical depths affects the calculated H₂O index values, while the phase function parameters do not. This observation is supported by Figure 6, which shows that spectral contrast is affected the most by the assumed τ and not the scattering behavior.

- Ice-rich surfaces (or spectra with strong band depths/greater spectral contrast) are affected more than 'dirtier' ice due to the nonlinear relationship of band depth and albedo in reflectance data. Therefore, assuming the incorrect τ propagates the most uncertainty for surfaces with high spectral contrast/deep absorptions.
- Assuming no aerosols causes the calculated H₂O index to deviate up to 0.08 (or 8%) relative to the control run.

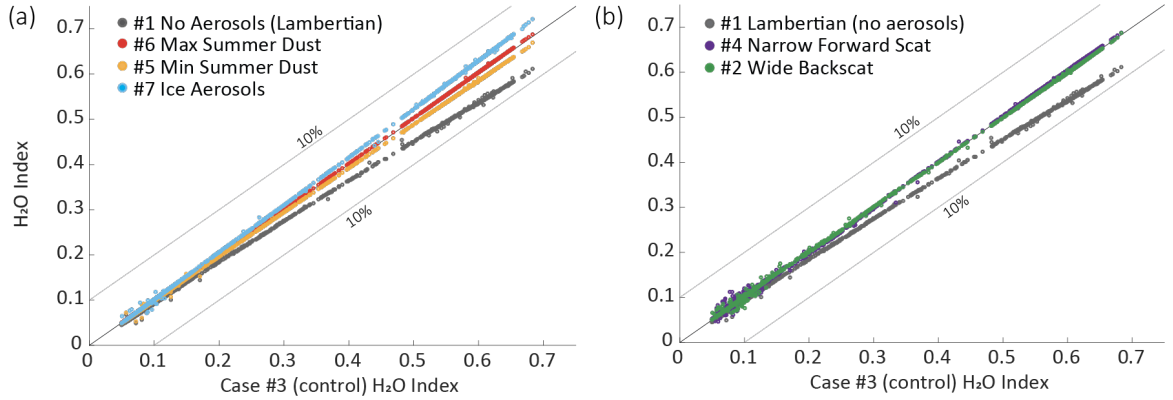


Figure 8. The plots above compare the H₂O ice band depth results between the different parameter cases. The scatter plots compare the model results from the control case - Table 1, Case #3 - (x-axis) to all the other cases (y-axis). This approach for case comparison was taken because ground truth / laboratory data, which would normally be used in this type of comparison, does not exist. Results that are close to the control case plot tightly to the one-to-one line (solid black line). The left plot compares the cases where τ was varied while the phase function parameters were held constant. The right plot compares the cases where phase function parameters were varied while τ was held constant.

4.4.1 Sensitivity of the H₂O index to optical depth uncertainty

Since the treatment of aerosols affects the relative absorption strengths, we would expect them to affect the calculated H₂O index values. Figure 8a has the H₂O index of each aerosol run plotted relative to the H₂O index from the control run (Table 1, Case #3). Uncertainty in knowing the absolute τ caused the greatest change in the H₂O index for more ice-rich spectra compared to dustier spectra. Leaving dust aerosols untreated causes ice absorptions to be weaker and the spectral slope to be redder. Not knowing the absolute τ at the time, day, and location that the data was acquired (i.e., using minimum to maximum summer dust opacity estimates from the MCD) led to a <2% difference in the H₂O index. Using $\tau_i = 0.05$ for one of the runs caused the H₂O index to increase by 3% relative to the control run. Leaving the aerosols untreated caused the largest decrease in the H₂O index by 8% for the most ice-rich spectra relative to the control run.

4.4.2 Sensitivity of the H₂O index to phase function parameter uncertainty

Band depths do not change when different phase function values are assumed. Figure 8b shows this quantitatively by plotting the H₂O index from the different retrievals against the H₂O index of the best SSA retrieval run (Table 1, Case #3). The H₂O index from the narrow forward scattering and modest backscattering runs are nearly identical to the modest forward scattering control run. The isotropic case's relative deviation in the H₂O index to the control run, as well as the other two scattering cases, is attributed to the isotropic case assuming τ is 0. The H₂O index's insensitivity to the different assumptions for the scattering variables, b and c , in the single-particle phase function is because they are constant as a function of wavelength, as was discussed in Section 4.2.2.

4.5 Differences in spectral results from using EPFs (DISORT EPF approach) versus the central scene (simplified approach).

Retrieving SSA through the DISORT-EPF approach is done using a wide range of phase angles (45°–120°) from the EPFs to match I/F to SSA. Performing SSA retrieval using all nine EPF observations creates a very large look-up table, thus increasing the computational expense. However, the tradeoff of using EPFs versus the central scene observation results in retrieving SSA values closer to the absolute/true SSA values, which decreases the uncertainty propagated into Hapke-modeled abundances and grain sizes. The most common and simpler method for converting I/F to SSA—when the DISORT routine is not an option for the scientist—is to use the phase angle coverage from the central CRISM scene, which only covers phase angles ~60–85°. This more common method we refer to as the "simplified approach" as described in Section 2.2.4. Due to the differences and rigor in the data processing and retrieval of SSA, we wanted to see how different the spectral results would be between these two approaches. For proper comparison, the common but incorrect assumptions that there are no aerosols to be treated and that the scattering behavior is isotropic were used as input parameters for both approaches.

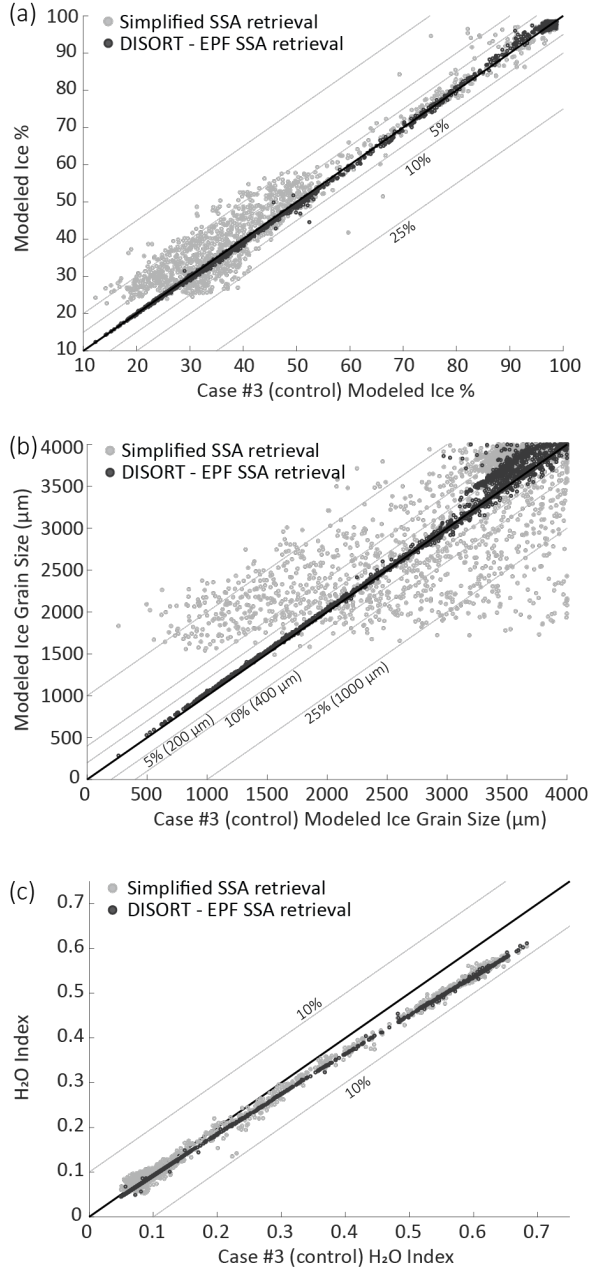


Figure 9. The plots above compare the Hapke unmixing (a, b), and the H₂O band depth (c) results between the isotropic-no aerosol DISORT case and the isotropic-no aerosols SSA retrieval that did not utilize the EPFs. The comparison with reference to the results from Case #3 (solid black line).

modeled ice abundance for pixels with higher lithic contents (up to 10–15% difference) (Figure 9a, light grey). The Hapke model prefers higher or lower ice abundance to produce the best spectral fit to the overall increase or decrease in spectral albedo induced by the two different SSA retrieval approaches. For example,

When we look back at Figure 6 and compare the two isotropic SSA retrievals' spectra, we see that the absorption shapes and depths are the same. What differs is the range in spectral albedo for the dirty, NPLD ice and the pure, NPRC ice between the two retrieval methods. The spectral albedo of the NPRC was 5%–0% darker from the simplified SSA retrieval compared to the DISORT-EPF SSA approach. However, the spectral albedo of the NPLD was brighter from the simplified SSA retrieval (by 2%) compared to the DISORT-EPF SSA approach. This range limit on the SSA retrieval between the brighter NPRC and darker NPLD surfaces is due to the limited phase angle range used in the simplified method.

The difference in the SSA retrievals discussed above had a profound effect on the Hapke-modeled ice abundance/grain size estimates between the two approaches (Figure 9a, 9b); the H₂O indices were nearly identical (Figure 9c). The simplified SSA retrieval resulted in a minimum modeled ice grain size of 1500 μm, while the minimum modeled ice grain size from the DISORT-EPF SSA approach was as small as 500 μm. In addition, there is an inconsistency in the

Figure 10 shows the Hapke forward modeled versus retrieved SSA spectra from the two different correction methods (FRT0000BEB8 in sensor space, pixel [380, 257]) and the corresponding abundance and grain size estimates. Despite the spectrum from the simplified approach leading to a small increase in SSA, by 1.5%, it was enough to force the Hapke model to increase the estimated abundance by 11% and the ice grain size by 1,045 μm while decreasing the abundance and grain size estimates of the lithic components. This minor difference in albedo shift appears to affect the trough wall's dirty-ice spectra, where the surface ice absorptions are much more subdued yet still present in the spectra.

5 Interpretation of sensitivity tests and suggestions for future CRISM studies

5.1 Hapke

Our seven different DISORT-EPF SSA assessments showed that the modeled grain size estimates were most sensitive to the over/under correction of ice aerosols, while the assigned phase function value affected the estimated ice abundances the most. However, it is safe to assume that ice aerosol opacity is low enough to be ignored for investigations using CRISM observations acquired between $L_s \sim 100\text{--}160^\circ$. The effects of dust opacity resulted in minor changes to the abundance or grain size estimates, implying the expected dust opacity range during northern summer from the Mars Climate Database are reasonable values to use for a given observation. To reasonably minimize error propagation, we suggest constraining the appropriate opacities (using the MCD) and scattering behavior (e.g., frost vs. regolith vs. ice) for the appropriate martian season, especially for studies doing time series, seasonal, or spatial-temporal analysis through spring to fall. Surfaces in spring and fall are prone to CO_2 frost-cover ($L_s < 100^\circ$ and $L_s > 164^\circ$) or fine-grained H_2O frost, which sublimates between $L_s = 93\text{--}110^\circ$ (Kieffer & Titus, 2001; Calvin & Titus, 2008), making phase function parameter assumptions more important due to its influence on abundance estimates. Hapke modeling already inherently has large error bars (10–30%) (Lapotre, Ehlmann, & Minson, 2017; Lapotre, Ehlmann, Minson, et al., 2017; Harris & Grindrod, 2018); therefore, all possible measures should be taken minimize these effects.

The uncertainty in knowing the atmospheric and surface properties influences the uncertainty of grain size estimates. Of the seven parameter cases, 95% of the estimated ice grain sizes deviated under 10% from the control. The deviation was greater for spectra modeled as ice-rich and larger grain size, likely due

to the model attempting to find the best fit that satisfies both the albedo and spectral contrast. This is especially challenging attempting to model highly opposing endmembers: ice—with its very high albedo and strong spectral contrast versus dust/palagonite—with is comparatively low albedo and featureless spectra.

This work shows studies should prioritize using EPFs for data correction and conversion because of the effects it has on quantitative analyses. Although this has been known, it was never quite clear how one's data processing methods would quantitatively affect the spectral analysis and subsequent interpretation, especially for CRISM polar data. In Section 4.5, we showed how retrieving SSA using EPFs, which covers a phase angle range of 45° – 120° , greatly impacts the surface properties' interpretation (i.e., Hapke nonlinear spectral unmixing results). Compared to the simpler method that uses only the central scene in the retrieval process (phase angle range of 60° – 85°), the abundance and grain size estimates vary considerably between the two processing and SSA retrieval approaches. Future polar ice studies with spectral modeling objectives should use atmospheric and photometric correction methods that use the widest phase angle range possible to retrieve reflectance/SSA of VNIR-SWIR hyperspectral data.

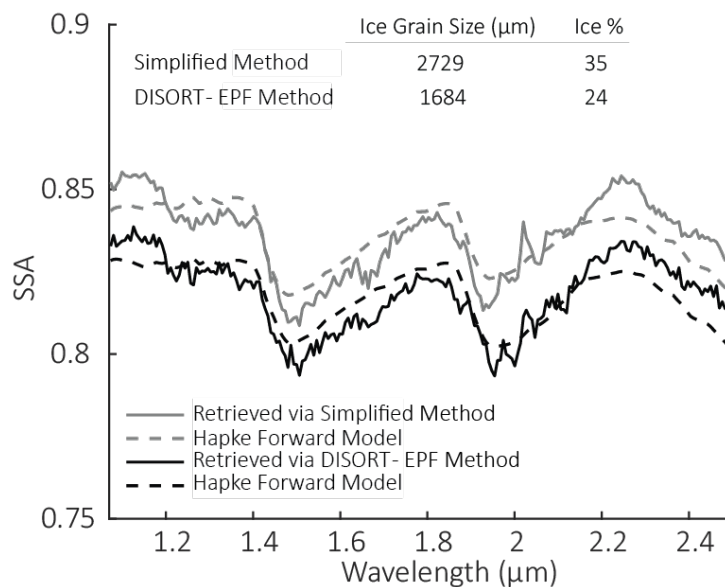


Figure 10. Plot of single-scattering albedo spectra from a single NPLD pixel retrieved from the DISORT-EPF approach (black) in comparison with the simplified retrieval (grey), which does not use the EPFs to obtain SSA (FRT0000BEB8 sensor space $[x, y]$ coordinates $[380, 257]$). Dashed lines are the best fit Hapke forward modeled spectra for the two spectra from the different approaches. The best fit modeled ice grain size and abundance are displayed above the spectra for convenience. The lithic components (dust/palagonite, respectively) had modeled grain sizes of $17\mu\text{m}/130\mu\text{m}$ and abundances of 5.3%/70.6% from the DISORT processed data while the simplified approach resulted in modeled estimates of $14\mu\text{m}/105\mu\text{m}$ and 2%/63% for grain size and abundance.

5.2 H₂O index

Band depth calculations are more sensitive to aerosols than phase function. Our results agree with the uncertainty analysis made by Brown et al. (2016); variation in the H₂O index <10% should not be interpreted as true surface changes when tracking the H₂O index seasonally across multiple observations. This uncertainty is greatest for extremely bright surfaces like ice and would likely apply to band depth parameters of bright mineral deposits elsewhere on Mars. Dust-covered surfaces that are not as high albedo or materials with weaker absorption bands are less affected. The H₂O index was nearly identical between the simplified SSA retrieval and the DISORT-EPF SSA retrieval (both assuming no aerosols and isotropic scattering). Future studies should not worry about fine-tuning the aerosol and surface properties if using the H₂O index as their main means of analysis. Therefore, if a study's approach plans to use the H₂O index or other band parameters to investigate seasonal surface change, a computationally simpler correction that does not utilize DISORT is acceptable if the 0.1 is used as the change detection limit. These results agree with the error analysis performed by Brown et al. (2016).

6 Conclusions

How does using a DISORT-EPF SSA routine (which utilizes the full viewing geometry of CRISM EPFs) change the spectral results compared to the more common and simplified SSA retrieval method that only uses I/F and viewing geometries from the near-nadir, central CRISM image?

- We found the biggest factor in SSA spectral sensitivity is whether the CRISM EPFs are used to retrieve SSA. This caused a large discrepancy in Hapke mixture modeling results between the more rigorous DISORT routine used here versus the simpler method that only employs the central CRISM scene. In other words, using the EPFs and DISORT to convert I/F to SSA produces profoundly different Hapke results compared to the simplified approach. However, future studies that do not need SSA spectra for their spectral analysis and instead only need I/F to band depth analysis do not need to worry about using EPFs.

How do untreated aerosols affect the spectral characteristics and interpretation of the ice-rich surface through typical analyses such as H₂O Index parameters and Hapke mixture modeling?

- The summer minimum dust optical depth, which was close to the opacity value derived from the Mars Climate Database (MCD) for the date and time of the observation, gave the best SSA retrievals (Section 4.1). The spectral analysis results were minimally impacted even when the data from the minimum and maximum summer dust opacities were compared. Therefore, using the MCD dust opacity estimates for the date, time, and location of a given observation is sufficient for data correction and analysis.
- Including ice aerosols or assuming an isotropic surface yields unrealistic SSA spectra (>1.0 or <0.0 SSA). Therefore, assuming the NPRC and NPLD are isotropic as it has been in the past is an inappropriate assumption. During the northern hemisphere summer, it is safe to assume ice aerosol optical depth is negligible.
- Not treating aerosols has the largest impact on H₂O reflectance band depths. The effect is greatest for the NPRC ($\Delta 8\%$). Its effect on lower albedo surfaces is minimal and can be ignored. However, accounting for seasonal variation in aerosol content would minimize error bars for temporal studies and should be corrected.

How does assuming isotropic versus anisotropic scattering affect spectral characteristics and interpretation of the ice-rich surface (i.e., Hapke mixture modeling)?

- First-order estimate of the phase function (b and c in $P(g)$) of the NPRC and NPLD: out of the four different scattering behaviors tested, both types of polar surfaces are best modeled as broad, moderate forward scatterers. This assumption is better than the traditional method of assuming isotropic scattering. Phase functions can be better constrained in future studies dedicated to the task. These studies should also investigate macroscopic roughness.
- Although the assumed surface phase function only effects the Hapke unmixing results (and only to a minor degree), accounting for the appropriate scattering behavior is needed to retrieve the most accurate SSA. Doing so decreases the propagation of uncertainty in estimated abundances and grain sizes, which is already inherently large due to the nature of assumptions going into nonlinear spectral mixture modeling (Section 3.1.1).

- Application to other worlds: although Europa and Enceladus do not have appreciable atmospheres, scattering behavior should be constrained before performing hyperspectral analysis and interpretation. Previous studies have constrained gross-estimates for macroscopic roughness and scattering direction and scattering lobe width for the different hemispheres of several icy worlds (e.g., Domingue et al., 1997); however, better spectrometers and imagers means needing to constrain different surfaces and their scattering properties based on their morphology and composition at the resolution or scale obtained by the imager. Macroscopic roughness—which was not explored in our study—will need to be constrained depending on the observations' resolution and ground sampling distance. This is something that future flyby missions like Europa Clipper should keep in mind when developing their data processing pipelines early on.

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Appendix A. Surface scattering model equations and non-parameterized constraints used in DISORT

The following lists the equations used to retrieve the average single scattering albedo.

- Surface scattering is defined according to the Hapke bidirectional reflectance equation (Hapke, 1981, Eqn. 16; Hapke, 1993, Eqn. 8.89), which is then converted to a radiance coefficient by multiplying r by π/μ_0 (Hapke, 1993, Eqn. 10.3–10.4) and multiplied by the shadowing function (Hapke 1993, 2012, Eqn. 12.54) (see Section 2.2.2):

$$r_c(i, e, g) = \frac{\omega_{ave}}{4} \frac{1}{\mu_0 + \mu} \{ [1 + B(g)]P(g) + H(\mu_0, \omega_{ave})H(\mu, \omega_{ave}) - 1 \} S(i, e, g)$$

where r_c is the “reflectance” or, in this case, I/F from the data and ω_{ave} is what is being solved for using the look-up table approach.

- $B(g)$ is the opposition effect and is defined by

$$B(g) = \frac{B_0}{1 + (1/h) \tan g (g/2)}$$

where B_0 is the amplitude which is correlated to the opacity of the surface particles (a measure of the percentage of incident light that is scattered at the surface; $B_0 = 1$ means all light is scattered, and the particle is opaque). However, the influence that B_0 has on the shape of the phase curve is only effective below phase angles of 30°.

- Lastly, h is the angular width parameter of the opposition effect and is approximated by

$$h = \frac{-3}{8} \ln(1 - \emptyset), \text{ where } \emptyset \text{ is the pore filling factor.}$$

- The H-function, describing multiple scattering of light, is approximated by

$$H(x, \omega_{ave}) = \frac{1}{1 - (1 - \gamma)x [r_0 + (1 - 0.5r_0 - r_0x) \ln(\frac{1+x}{x})]}$$

where x is either μ_0 or μ , $\gamma = \sqrt{1 - \omega_{ave}}$, and $r_0 = \frac{1-\gamma}{1+\gamma}$.

- The shadowing function (Hapke, 2012, Eqn. 12.52-12.54) accounts for macroscale roughness, $\bar{\theta}$, characterized by a mean slope angle for cases when $e < i$ and is expressed by the following:

$$S(i, e, \psi) \simeq \frac{\mu_e}{\eta_e(e)} \frac{\mu_0}{\eta_{0e}(i)} \frac{\mathcal{X}(\bar{\theta})}{1 + f(\psi)\mathcal{X}(\bar{\theta})[\mu/\eta_e(e)]}$$

where ψ is the azimuthal angle made between the azimuthal planes made by the CRISM detector and sun, which is calculated from i, e , and g . $f(\psi)$ is the fraction of the visibility shadow that is hidden in the shadow cast by an “object.” The terms μ_e , μ_0 , and $\mathcal{X}(\bar{\theta})$ help describe the effective tilt of the surface. $\eta_{0e}(i)$ and $\eta_e(e)$ describe the effective cosines when the azimuthal angle, ψ , is zero. The lengthy and mathematically complex expressions and derivations for the above terms within the shadowing function can be found in chapter 12 of Hapke 2012 or Hapke 1993.

Table A1. Non-parameterized constraints used in DISORT

DISORT	
<i>Dust aerosol grain radii</i>	1.5 μm (Wolff et al., 2009)
<i>Ice aerosol grain radii</i>	2 μm (R. Todd Clancy et al., 2003; Wolff & Clancy, 2003)
<i>Dust aerosol distribution</i>	Constant vertical mixing ratio, 0–150 km altitude
<i>Water ice aerosol distribution</i>	Well mixed, >15 km altitude
<i>Roughness $\bar{\theta}$ (assumes Gaussian slope distribution)</i>	9°

Appendix B. Equations used for Inverse Hapke Modeling of the Retrieved SSA Data Using the Optical Constants of Mixture End-members

The following equations are from Hapke (2005) and were used in the application of the Newton's method to the inverse RTM code developed by Li and Li (2011), which was modified for this study.

- Scattering efficiency of i th (ice, dust, or palagonite) component:

$$Q_{si} = S_e + (1 - S_e) \frac{(1 - S_i)\theta}{1 - S_i\theta}$$

- Fresnel reflectivity for externally incident light (n & k are the real and imaginary parts of the complex refractive indices):

- $S_e = \frac{(n-1)^2 + k^2}{(n+1)^2 + k^2} + 0.05$

- Fresnel reflectivity for externally incident light:

$$S_i = 1.014 - \frac{4}{n(n+1)^2}$$

- Internal transmission coefficient of the particle without internal scatters for irregular particles (Shkuratov & Grynko, 2005) (D is the particle diameter):

$$\theta = e^{-\alpha 0.2D}, \text{ where } \alpha \text{ is the absorption coefficient for the internal attenuation: } \alpha = \frac{4\pi nk}{\lambda}. \text{ Note that}$$

if the empirical relationship of grain size and optical pathlength were assumed to be spherical (as under Hapke's model), the estimated grain size diameter, D , would be $\sim 1/3$ to $1/4$ of the irregular grain diameter estimates. Which is the better assumption for icy and dusty icy surfaces has yet to be investigated. We simply assumed palagonite, dust, and ice particles were irregularly shaped since the purpose of this study was the relative change in modeled grain size.

- The calculated single scattering albedo of the individual endmembers can then be added linearly to model the average single-scattering albedo for an intimate mixture,

$$w_{ave} = \sum_i \frac{M_i Q_{si}}{\rho_i d_i} / \sum_i \frac{M_i Q_{se}}{\rho_i d_i}$$

Table A2. Non-parameterized assumptions and variable ranges used in the Hapke spectral mixture modeling

Hapke		Comments
<i>Opposition surge amplitude, B_0</i>	1.0	Optical constants: (Roush, 1993; R. T. Clancy et al., 1995) Particle density: could not be found for the palagonite samples measured by Roush. Density of altered Icelandic palagonite was used instead (Franzson et al., 2010)
<i>Fill factor, Φ</i>	0.41 (porosity 59%)	
<i>Palagonite particle density</i>	2.65 g cm ⁻³	
<i>Dust particle density</i>	1.91 g cm ⁻³	Optical constants: (Wolff et al., 2006, 2009) Particle density: does not exist for the OC derived from atmospheric dust so the mean density of “volcanic ash” JSC Mars-1 (Allen, Jager, et al., 1998; Allen, Morris, et al., 1998) was used.
<i>Water ice solid density</i>	0.9167 g cm ⁻³	Optical constants from (Warren & Brandt, 2008)
<i>Particle shape</i>	irregular	
<i>Particle diameter range</i>	0 – 4000 μ m	

Supplementary Materials

S1. Determining the best SSA retrieval

Our results from testing the three different phase function parameter cases (main text Table 1, Case. #2–4), in which aerosol optical depth was held constant, suggest that assuming broad-moderately forward scattering behavior yields the lowest mean Δ SSA (Table 1, Case #3) for both the NPRC and trough wall. This result implies both surfaces are best modeled generally as broad, moderate forward scatterers. Such scattering behavior is typical for particle aggregates and/or materials with moderate internal structure (A. F. McGuire & Hapke, 1995), which is consistent with what we would expect for imperfect ice crystals, dirty-ice, and dust aggregates.

Next, the broad-moderately forward scattering behavior was held constant in order to assess which aerosol optical depths best describe the state of the atmosphere at the time the data were acquired (Table 1, Cases #3 and #5-7). Using the lowest mean Δ SSA method, our results suggest that using the approximated minimum summer dust optical depth yields better SSA retrievals than the control MCD τ_d estimate. Overall, we found broad-moderately forward scattering, lower τ_d , and 0 τ_i provides the best SSA retrievals out of the seven other cases.

Out of the seven different cases tested, three of them yielded unrealistic $SSA(\lambda, g)$ values. Case #4 (narrow forward scatterer), #7 ($\tau_i = 0.05$), and #1 (isotropic, no aerosols treated) resulted in the largest mean Δ SSA values of up to 0.15–0.2 Δ SSA (main text Figure 5). For comparison, the best case, #5, returned a mean Δ SSA value of <0.1 . Therefore, the SSA-phase retrieval curves for cases #1, 4, and 7 deviated away the most from having constant $SSA(\lambda)$ regardless of the viewing geometry. Additionally, these three cases resulted in failed SSA retrievals (Figure S2, S3, S4). A failed retrieval is when any retrieved SSA value is >1.0 or ≤ 0.0 for any phase angle range.

When a narrow forward surface scatterer (with “control” τ_d , $\tau_i = 0$) or isotropic scatterer (no aerosols treated) was assumed for the NPRC, at high phase angles, the SSA retrievals were >1.0 across the visible wavelengths (0.7–0.9 μm). Therefore, both a narrow forward scatterer and diffuse reflector are unrealistic scattering assumptions for the NPRC surface at visible wavelengths. Lastly, assuming a modest τ_i of 0.05 instead of 0.0 results in SSA retrievals bottoming out at 0.0 SSA at the wavelength regions for water ice

absorptions (1.5, 2.0, or roll-off past $\sim 2.4 \mu\text{m}$). In this case, when we assume ice aerosols are present, the model attempts to retrieve negative SSA values at those deep, water ice absorptions due to the overcorrections of water ice aerosols. Therefore, it is best to assume that negligible ice aerosols are present for this particular CRISM scene.

Table S1. Pixel resolution of EPFs and number of pixels covered by the “SSA assessment” ROIs

EPF observation	01	02	03	04	05	07	09	0a	0b	0c	0d
<i>Resolution (m/pixel)</i>	463.2	375.4	325.3	286.3	263	21.3	271.1	300.4	334.4	394.3	473.9
<i>Trough ROI size (pixels)</i>	30	41	51	61	77	8,306	70	59	46	39	26
<i>NPRC ROI size (pixels)</i>	38	52	70	79	101	11,455	86	74	61	49	35

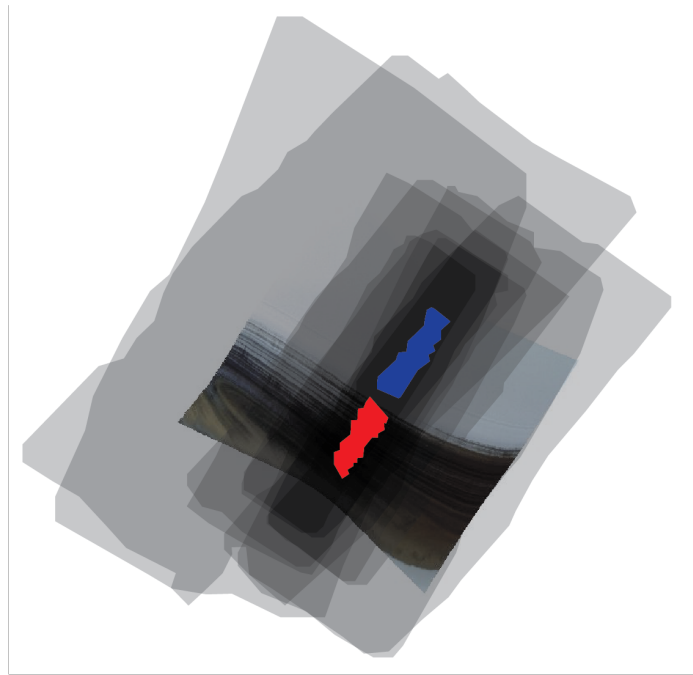


Figure S1. Location of the ROIs (blue = NPRC, red = trough wall/NPLD) used in the best SSA retrieval determination. ROIs were drawn over surface areas that overlapped across all EPFs. The overlapping areas of the EPF observations (gray footprints with transparency set to 25%) with the central “butterfly” (07) scene and ROIs are depicted above. Footprints of EPF observations are displayed in gray with their transparency set to 25%. The region with 100% overlap from all observations is the region with the darkest shade of gray. The number of pixels of each ROI differ for each EPF due to the EPFs having variable pixel resolution (due to spacecraft distance and observation mode). See Table S1 for EPF resolutions and corresponding ROI size in number of pixels.

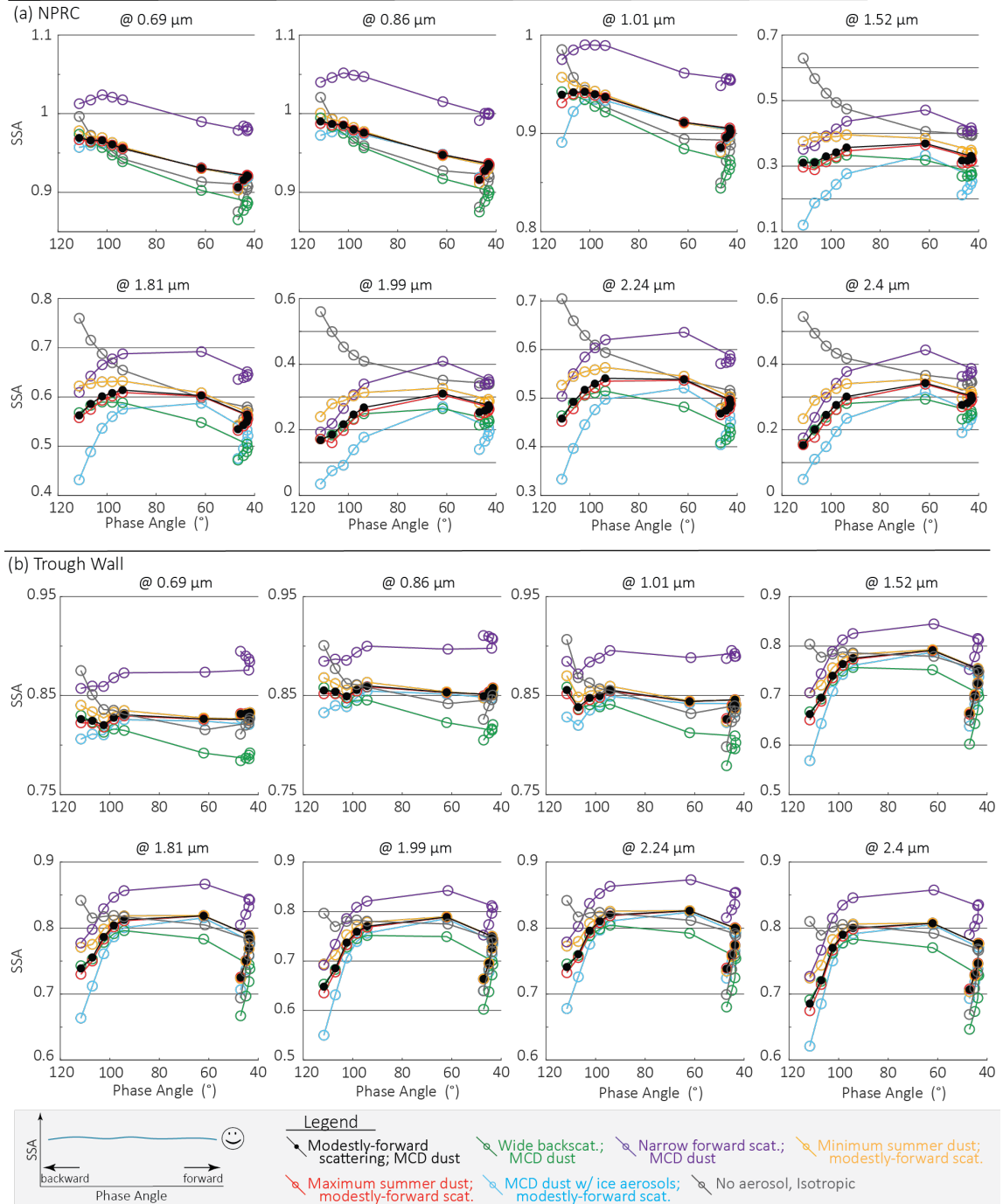


Figure S2. Examples of SSA-phase retrieval curves at select wavelengths. SSA-phase retrieval curves are another means of assessing how SSA was retrieved as a function of phase angle and wavelength. The flatter the curve, the better the retrieval since, by definition, SSA should be independent of viewing geometry. As mentioned in Section 2.2.3 of the main text, data points from the SSA-phase retrieval curves were used to calculate Δ SSA (main text Figure 5). See Figures S3 and S4 for a different visualization of viewing and assessing the SSA retrievals.

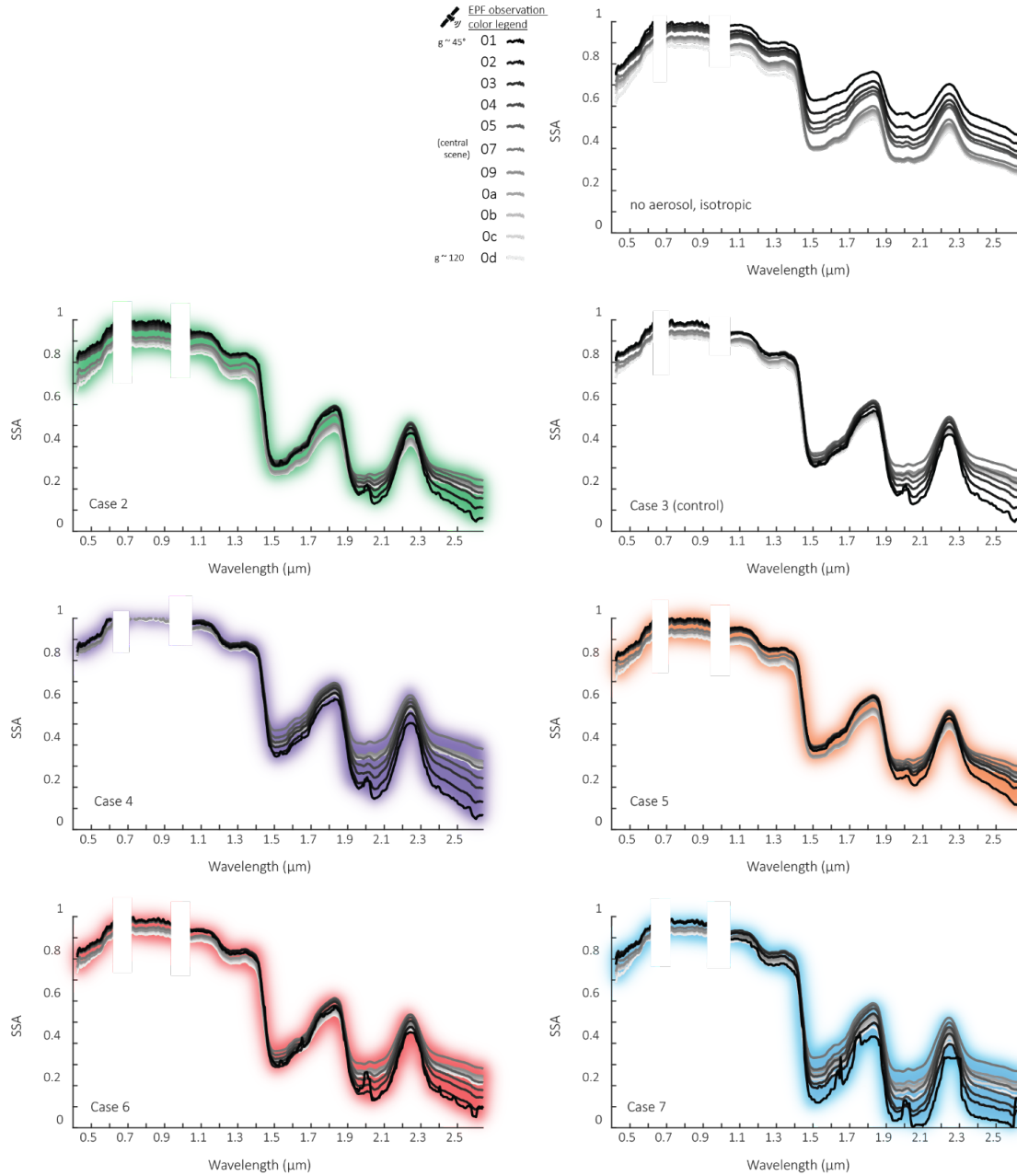


Figure S3. Average SSA spectra from the blue ROI (NPRC) in Figure S1 from each EPF observation. Perfect SSA retrievals would show the spectra plotted overlapping one another perfectly. Grayscale legend corresponding to EPF observations: darkest $\rightarrow$ lightest = EPF 01 (phase angle $\sim 45^\circ$) $\rightarrow$ EPF 0d (phase angle $\sim 120^\circ$). Colored shading behind spectra is a visual aid correspond to the Case #.

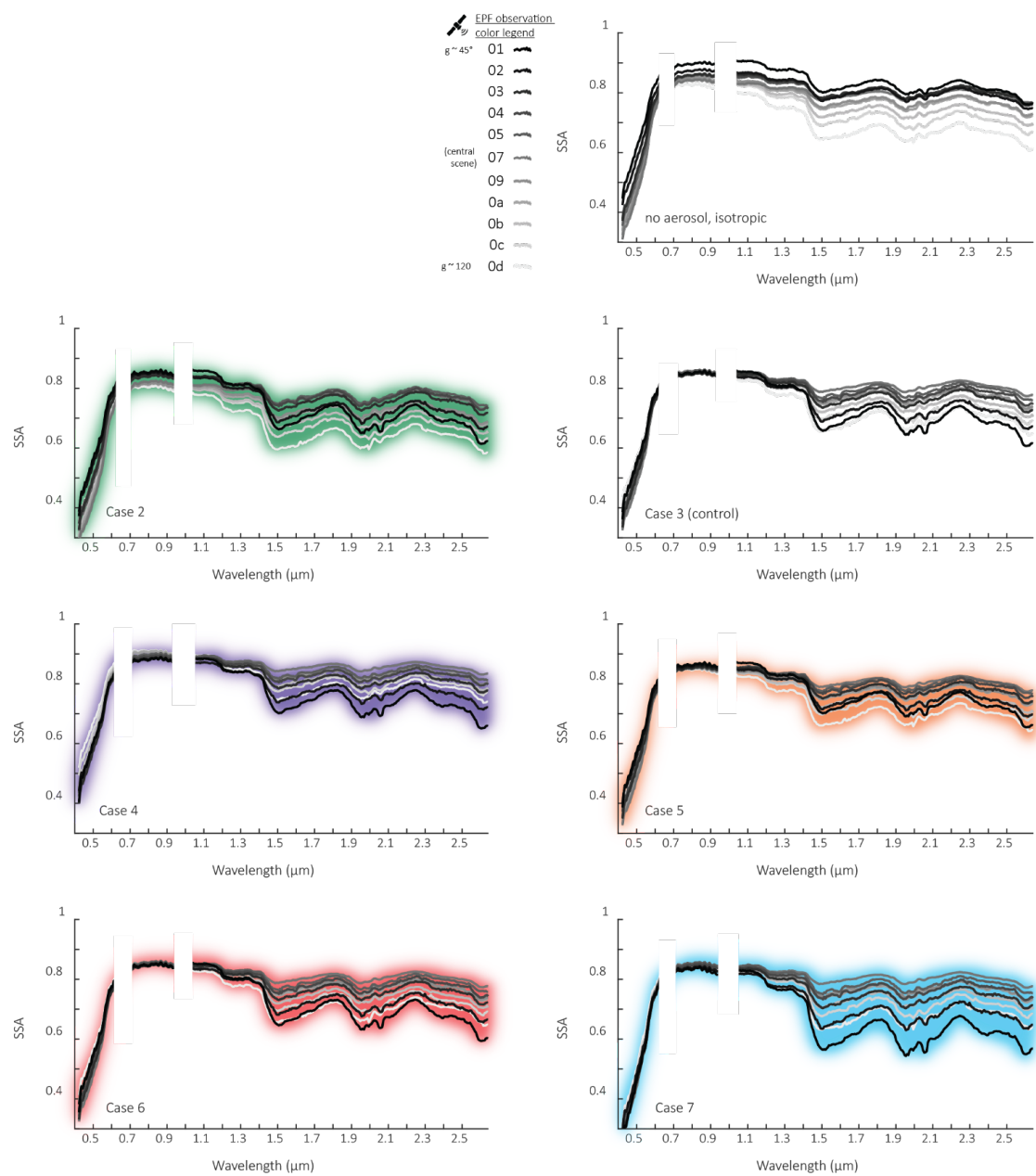


Figure S4. Average SSA spectra from the red ROI (trough wall/NPLD) in Figure S1 from each EPF observation. Perfect SSA retrievals would show the spectra plotted overlapping one another perfectly. Grayscale legend corresponding to EPF observations: darkest $\rightarrow$ lightest = EPF 01 (phase angle $\sim 45^\circ$) $\rightarrow$ EPF 0d (phase angle $\sim 120^\circ$). Colored shading behind spectra is a visual aid correspond to the Case #.

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Chapter 4

Present-day surface ice/lithics properties of Mars' North Polar Layered Deposits controlled by recent post-depositional processes and regional depositional variability

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Abstract

The lithic-to-ice ratio of layers in the ice-rich North Polar Layered Deposit (NPLD)—particularly of marker bed (MB) layers—are believed to hold the key to unraveling our understanding of the recent Amazonian climate record. Despite decades of investigations with visible and ground-penetrating radar data, the compositional variability between the layers and their connection to climate-driven accumulation processes remains ambiguous due to the superposition of modern aeolian surface processes (i.e., trough wall erosion, sublimation, redistribution of dust and ice, and dust-veneer formation and induration), which modify the surface composition and topography. This has discouraged researchers from using visible to near-infrared (VNIR) spectral data to characterize the stratigraphy. However, correlating spectrally derived surface composition with spatially resolved layering can aid in deconvolving superimposed processes from primary accumulation processes. We rigorously evaluate the MB ice/lithic abundance and grain sizes in the NPLD to determine MB variability within and between two well-exposed trough sites using Hapke nonlinear spectral unmixing of CRISM data to determine whether the surface properties of MBs are primarily a function of past environmental conditions during layer accumulation or recent post-depositional aeolian processes. We found that the spectral derived layer properties are inherently a mixture of variability in local paleo-conditions during MB accumulation and the variability in modern-day post-depositional processes. More conclusively, our results suggest that MBs with ice abundances >60% and that are statistically similar between multiple dust veneer-free troughs sections can most confidently be linked to initial layer properties (i.e., MB3). We also found that veneer-free sections of the investigated troughs (specifically strata deeper within the trough) are still affected by superimposed modern aeolian processes specifically to strata deeper within the trough due to winds and preferential collection of eroded debris downslope. Lastly, our results suggest that dust veneers that completely blanket NPLD trough wall exposures are optically thick and have a consistent ice-to-lithic ratio (~35% lithics) implying it is likely a lithic deposit cemented together by pore-filling ice. These results confirm Rodriguez et al. (2007)'s hypothesis that veneers are composed of sand-sized particulates held together by pore-ice. Future work looking to integrate radar, elevation, and hyperspectral data could further constrain the MB formation processes and their insight into the climatic history of the Amazonian.

1 Introduction

Since its discovery in the 1970s, the ~1500 m thick dome of distinctly layered ice, known as the north polar layered deposits (NPLD), has been a target of investigation due to its putative record of current and recent past martian climate history. It is hypothesized that the NPLD's bright and dark-toned layers (e.g., marker beds), which are visible within a series of spiral troughs dissecting the north polar cap (Howard et al., 1982; Tanaka K. L. et al., 2003; Tanaka, 2005) record variations in relative abundances of water-ice and atmospherically deposited dust over the past <4 Myr due to cyclical changes in solar insolation (Murray et al., 1972; Cutts and Lewis, 1982; Laskar et al., 2002; Levrard et al., 2007; Perron and Huybers, 2009; Becerra et al., 2017). However, despite decades of investigations, the compositional variability between the layers and their connection to climate-driven accumulation processes remain ambiguous.

Earlier studies revealed a variety of surface textures and erosional styles throughout the vertical NPLD sequences: dark-toned, protruding knobby, and pitted marker beds (MB), rough or smooth undifferentiated and recessive beds, and thinly layered sets with variable resistance (Milkovich and Head, 2006; Fishbaugh and Hvidberg, 2006; Herkenhoff et al., 2007; Fishbaugh et al., 2010a, 2010b). These studies suggested that the erosional morphology, apparent brightness, and protruding nature of a layer (or set of layers) is a function of the ice-to-dust ratio driven by the insolation cycle. Other studies of the NPLD (Herkenhoff et al., 2007; Fishbaugh et al., 2010a) indicated that the morphology and brightness of the layering could be a function of seasonal wind and climate patterns controlled by trough topography. A clear link between composition and layer type, specifically among MBs, and factors influenced by trough wall geometry such as allochthonous dust veneers has not been demonstrated.

The interpretation and expression of climate-related primary depositional processes in the stratigraphy are complicated by the superimposition of modern aeolian surface processes (i.e., trough wall erosion, sublimation, redistribution of dust and ice, and dust-veneer formation and induration), which modify the surface composition and texture (Rodriguez et al., 2007; Fishbaugh et al., 2010a; Herkenhoff et al., 2007; Chapter 2 of this dissertation). These superimposed processes have discouraged researchers from using visible to near-infrared (VNIR) spectral data to characterize the stratigraphy. They have instead used VNIR data to study temporal variations of the topmost residual ice layer (i.e., north polar residual cap—NPRC) (Langevin et al., 2005; Calvin et al., 2009; Brown et al., 2012, 2016) or have focused on shallow radar

sounding (SHARAD (Seu et al., 2004); 10 m vertical resolution in pure ice) to investigate subsurface NPLD structure and internal composition (Phillips et al., 2008; Lalic and Holt, 2017; Lalic et al., 2019). The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) full-resolution targets (FRT) observations (18 m/pixel) (Murchie et al., 2007) have the untapped potential to retrieve high spatial resolution and temporal information about the surface composition of the stratigraphic layers.

Correlating spectrally interpreted composition with spatially resolved layering can aid in deconvolving the superimposed processes and constrain the accumulation history. As interpreted from radar data analysis, dust content ranges from 2-60% for the strong internal reflectors (Picardi, 2005; Phillips et al., 2008; Lalic and Holt, 2017; Lalic et al., 2019). Presently, the ratios of water ice to lithics (the majority of which is likely atmospherically deposited dust) of the NPLD layers, specifically the marker beds (Malin and Edgett, 2001; Milkovich and Head, 2005; Fishbaugh and Hvidberg, 2006; Fishbaugh et al., 2010a, 2010b; Becerra et al., 2016), have yet to be made from VNIR data despite having equal to or better vertical resolution than SHARAD (dependent on the surface slope and the angle at which they are exposed). Furthermore, current research has not established a genetic link between a layer observed at the surface and a strong internal reflector from radar in the subsurface (Christian et al., 2013). Despite this challenge, establishing compositional estimates through VNIR data would help bridge the gap between previous geomorphic and radar analyses by connecting the two with patterns in ice-dust ratio variability. This paper takes a crucial first step for such an approach: an initial investigation of marker bed composition and analysis of the main influences on their surface expression, be they post-depositional processes or intrinsic layer properties.

Interpreting the record in the stratigraphy is a two-step process that involves: 1) an accurate description of the stratigraphy by integrating and analyzing multiple datasets (i.e., radar, visible, hyperspectral); and 2) use of the stratigraphic record to constrain climate and layer accumulation model inputs for stratigraphic reconstruction (Levrard et al., 2007; Hvidberg et al., 2012; Hvidberg, 2016) Unfortunately, these accumulation models fail to model the observed layers properly. This implies that our current interpretation of the stratigraphy is incomplete due to the unconstrained NPLD composition and variability. Detangling the expression of present-day processes from the intrinsic layer properties is the next step needed to characterize the stratigraphy more completely. Motivated by previous studies, we rigorously evaluate the MB ice/lithic abundance and grain size in the NPLD to determine MB variability within and between troughs.

This data drives our discussion regarding the relationship between the surface expression of layers and surface composition: whether it is primarily a function of past environmental conditions during layer accumulation (Milkovich and Head, 2006) or recent post-depositional aeolian processes (Fishbaugh et al., 2010a). With these goals in mind, the scientific questions this study addresses are as follows:

1. What are the lithic and ice abundance and grain sizes (i.e., surface properties) measured by VNIR spectroscopic techniques of the NPLD?
2. Do VNIR constrained surface properties of regionally continuous MBs significantly vary between well-exposed NPLD outcrops?
3. Based on the regional variability, do VNIR constrained surface properties reflect the intrinsic properties of MBs (i.e., properties that represent original MB accumulation processes)?
4. How do the VNIR abundance estimates compare to the radar estimates?

2 Methods

2.1 Data and study locations

Polar CRISM observations allow for spectral measurements along the layered, gently sloping trough walls. The slope of the walls (3° – 20°) combined with the 18 m/pixel CRISM resolution translates to obtaining surface spectral information from layers that are as thin as 1–3 m in a vertical cross-section. We present an analysis of two VNIR hyperspectral image cubes acquired by CRISM: FRT0000BEB8 and FRT00003127, which cover two trough sites: N0 and N8, respectively (Becerra et al., 2016). Our model uses TRR3 L detector data between 1.0–2.5 μm . Metadata information for these observations can be found in Table 1. These two scenes were selected for several reasons: 1) overlap with HiRISE images and DTMs (needed for georeferencing, layer identification, and mapping), 2) date of data acquisition during optimal instrument operation temperatures and martian northern summer (to avoid seasonal CO_2 frost and dusty atmospheric conditions, clouds, and hazes (between $L_s = 95^{\circ}$ – 150°)), 3) a given set of layers in sequence are deemed to be regionally consistent between the two locations (Becerra et al., 2016); lastly 4) each location has unique spatial variability with regards to dust cover and trough orientation (slope/aspect) (Figure 1). More specifically, site N8 has well exposed NPLD layers with minimal dust veneer cover, while N0 is partially veneer-free and partially covered by a dust veneer.

Data processing and analysis was a three-part process: 1) atmospheric correction and retrieval of surface single-scattering albedo (SSA) using a radiative transfer model, 2) nonlinear least squares spectral unmixing of the single scattering albedo spectra, and 3) statistical assessment using multiple Student's t-tests of modeled abundances and grain sizes.

Table 1. Observation information

CRISM ID	FRT0000BEB8	FRT00003127
<i>L_s (°) (solar longitude)</i>	108.27	138.19
<i>Mars Year (MY)</i>	29	28
<i>Center Latitude & Longitude</i>	87.047 °N, 96.17 °E	86.536°N, 79.953 °E
<i>Local Solar Time (hour angle °)</i>	11.75	12.18
<i>UTC</i>	Aug 05, 2008	Nov 20, 2006

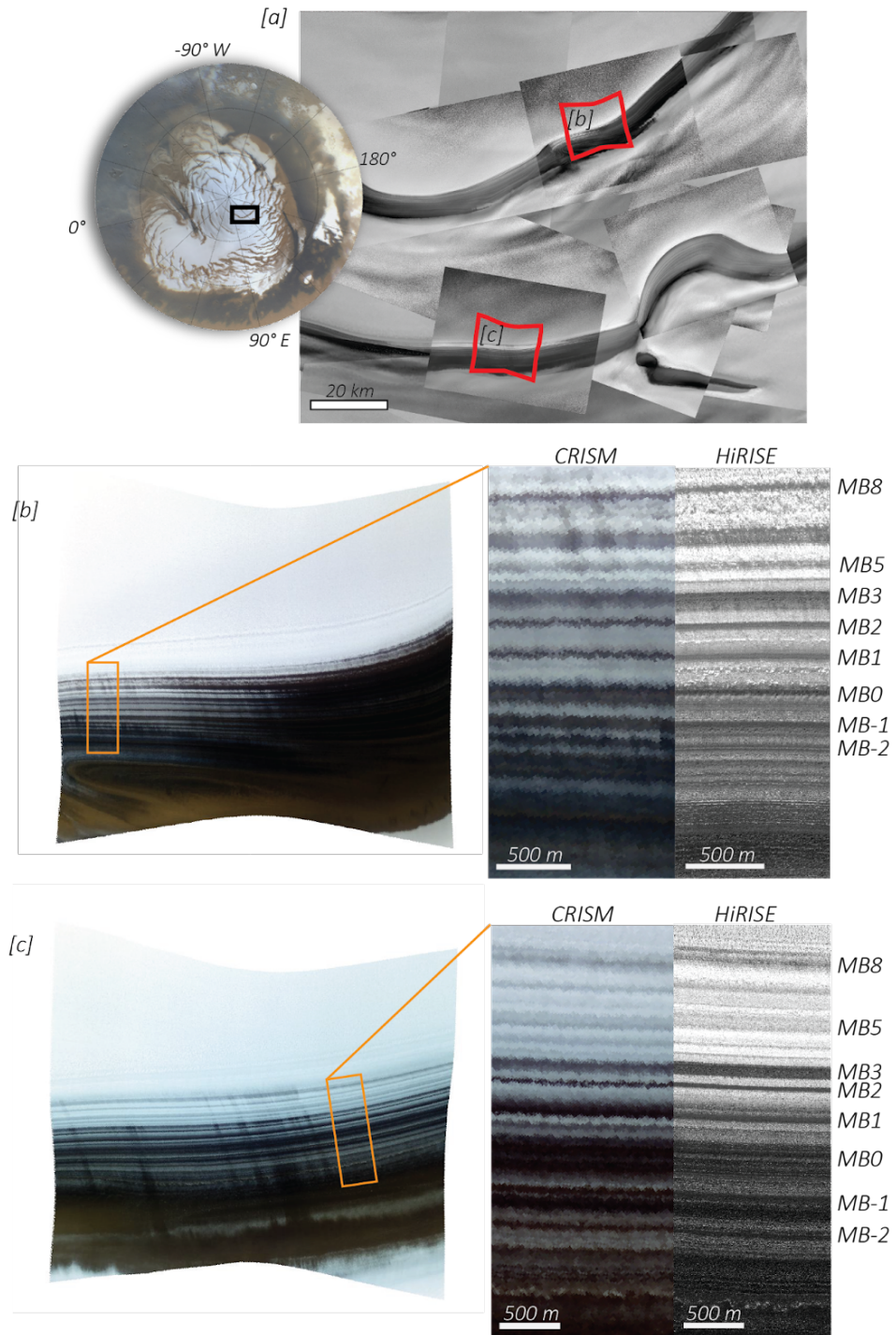


Figure 1. [a] Context image showing sites N0 and N8, the CRISM footprint locations at each respective site, all shown on CTX (Malin et al., 2007) mosaic basemap (Tanaka and Fortezzo, 2012). [b] and [c] show the CRISM data in true color (572.8 nm, 540.2 nm, 436.1 nm) accompanied by zoomed-in subset images of the NPLD stratigraphy for sites N0 and N8, respectively. For context and stratigraphic detail, the corresponding zoomed-in locations in HiRISE are shown to the right of the CRISM subset.

2.2 Atmospheric and photometric correction

The atmospheric and photometric correction was a two-step process. First, atmospheric gas bands from CO₂ were removed using the volcano-scan method (McGuire et al., 2009) with the CAT (CRISM Analysis Tool; available on the PDS) add-in toolbox in ENVI. Then a Discrete Ordinate Radiative Transfer (DISORT)-based atmospheric correction routine, developed at Washington University in St. Louis (Arvidson et al., 2006; Pascuzzo et al., 2021 (*in review/Chapter3*)), was used to separate the atmospheric aerosols and surface radiance contributions from the CRISM volcano scan corrected I/F (radiance at the sensor (I) divided by the incident solar irradiance divided by π (F)) spectra and retrieved single-scattering albedo (SSA). This pipeline uses the central CRISM scene and the EPF (Emission Phase Function) observations to retrieve more accurate SSA. A detailed description of the model and pipeline used can be found in Chapter 3 Section 2.2 and Chapter 3 Appendix (Pascuzzo et al., 2021, *in review*).

2.3 Nonlinear spectral unmixing

We employ a nonlinear Hapke radiative transfer model (Hapke, 1993) developed by previous studies (Li and Li, 2011; Li and Milliken, 2015; Robertson et al., 2016). The SSA image cube (observation 07—the central swath) was used to solve for abundance and grain sizes using an intimate ternary mixture model of water-ice (Warren and Brandt, 2008), martian atmospheric dust (Wolff et al., 2006, 2009), and palagonite (Roush, 1993) using their optical constants (Figure S1). Both palagonite and atmospheric dust are employed as endmembers due to the unknown nature of the lithics combined with the spectral variability of these two endmembers allowing for a better overall fit. Palagonite (i.e., altered basalt) was chosen as an endmember instead of unaltered basalt because its closer spectrally and compositionally to the expected lithic components that would be delivered through airfall. We use the following Hapke equation for intimate mixture modeling of relative component abundance and effective grain diameter (i.e., optical path length),

$$\omega_{ave} = \sum_i \frac{M_i Q_{si}}{\rho_i d_i} / \sum_i \frac{M_i Q_{se}}{\rho_i d_i}$$

, in which ω_{ave} is the average SSA, M_i is the mass fraction of component i , ρ_i is the solid particle density, d_i is the average effective particle diameter, and Q_{si} is the scattering efficiency of the i th component. In our

model, grain diameter is a free parameter given a selection range of 0–4000 μm . The above equation is solved for the mass fraction (or relative abundance in the case of orbital data) and the effective grain size of each i th component by forward modeling. When solving the nonlinear Hapke equations, numerous mass fraction and grain size combinations are tested (2000 iterations for each spectrum/pixel) until the best spectral fit is achieved using Newton's theory and Levenberg-Marquard least-square nonnegative algorithm for optimization (using the *lsqnonlin* function in MATLAB) (Li and Li, 2011; Li and Milliken, 2015; Robertson et al., 2016). The model outputs abundance and grain size maps for each component and a spectral cube of the forward modeled SSA spectra with an RMSE (root mean square error) band comparing the forward modeled spectral fit to the observed SSA data cube. For the complete set of equations used and parameterization of the unmixing, see Chapter 3 Section 3.1 and Chapter 3 Appendix B (Pascuzzo et al., 2021, *in review*).

Our model cannot currently allow differing assumptions in grain shape for different components. However, assuming a spherical shape for ice grains and an irregular shape for lithic components would be most geologically accurate. We, therefore, ran two unmixing models: one where all components are assumed spherical and one where all components are assumed irregular. Here we report ice grain size from the spherical run and palagonite and dust grain sizes from the irregular run. The irregular and spherical runs have a difference of approximately 6% for the ice vs. lithic abundances. Therefore, in order to keep consistent results and abide by our sum-to-one constraint, we report all abundances from our spherical run.

The modeled data cubes were georeferenced via tie points to a Context Camera (CTX) images (Malin et al., 2007) acquired during the same Mars year and solar longitude (L_s). To ensure the best georeferencing possible we spatially downsampled the CTX images to CRISM spatial resolution using pixel aggregation in ENVI then compared the resampled CTX image to the original CRISM S detector 0.600 μm band to match tonal brightness patterns, allowing for a more confident ground control point (GCP) selection between the two different datasets. GCPs made for the S detector data were applied to the SSA and model result image cubes. Different people did the georeferencing three times to assess pixel distance offset due to variability in GCP selection. Pixel offset across the georeferencing attempts was ~ 1 CRISM pixel near the center of the image and ~ 3 CRISM pixels (~ 60 m) towards the edges. After georeferencing the NPLD, exposure regions of interest (ROIs) were drawn for each marker bed shared between site N0 and N8: MB8,

5, 3, 2, 1, 0, -1, -2 (Figure 2). The data for each ROI was extracted and run through multiple Student's t-tests to assess composition and grain size variability for layers within the same site and between the two sites. The means and standard deviations for endmember abundances and effective grain sizes can be found in the Supplementary Material (Tables S1–S3).

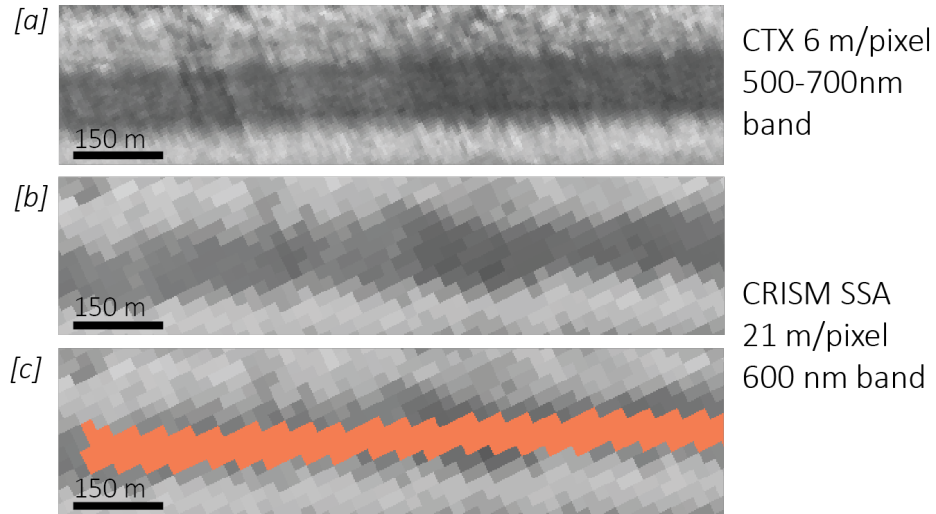


Figure 2. Example subset of georeferenced [a] CTX and [b] CRISM image (FRT0000BEB8) used to guide the pixel selection of MB ROIs. [c] Example of MB ROI overlain on the CRISM image shown in [b].

2.4 Statistical analysis

We use multiple t-tests to assess the statistical significance of the differences between N0 and N8 and within site N0 to compare veneer and veneer-free sections of the same layers. The unmixing results from each pixel produce six values: ice abundance, palagonite abundance, dust abundance, ice grain size, palagonite grain size, and dust grain size. Because the true make-up of the lithic components in the NPLD is unknown, and the lithic endmembers were chosen due to their spectral variability, we combine palagonite and dust abundance as "lithic abundance" to remain consistent with previous literature that does not separate the composition of the lithic component. Consequently, the lithic abundance is always the complementary percentage of the ice abundance due to this grouping and the sum-to-one constraint in our unmixing model. Thus, we condense our six results to four (ice abundance, ice grain size, palagonite grain size, and dust grain size) for our statistical analysis.

For the analysis within site N0, we compare a veneer covered and veneer free section for eight marker beds and a selection of interbeds averaged as one additional pair of ROIs. For the analysis between N0 and N8, we compare the veneer-free ROIs from N0 to the ROIs from N8 (which does not have a substantial veneer) for the eight correlated marker beds. With four unmixing values per comparison, this results in 68 total t-tests.

We use a two-tailed Student's t-test to determine if the means of two ROIs are statistically significantly different, implying that different geological processes control the surface expression of the two regions. The two-tailed test does not give preference to a difference in either direction of the means (one being statistically smaller vs. larger than the other); it simply tests if they are not the same. We use means weighted by the RMSE values of the unmixing runs in our t-tests; this decreases the effects from outlier pixels with poor fits while still including their results to an extent appropriate based on their RMSE. The equation for our t-test calculation is below,

$$t = \frac{|\overline{X}_1 - \overline{X}_2|}{\sigma_{pooled} \sqrt{\frac{n_1 * n_2}{n_1 + n_2}}}, \sigma_{pooled} = \sqrt{\frac{\sigma_1^2(n_1 + 1) + \sigma_2^2 * (n_2 - 1)}{n_1 + n_2 - 2}}$$

Wherein t is the calculated Student's t value, x is the weighted mean of each ROI, σ is the standard deviation (the equation for pooled standard deviation is also shown), and n is the number of pixels for each ROI.

The calculated t value is compared to the t value from the Student's t cumulative distribution function for the given degrees of freedom (total number of pixels from both ROIs minus 2), which produces a p -value. A p -value less than or equal to 0.05 suggests the two means are statistically significantly different within a 95% confidence interval. The t -test results can be found in the Supplementary Material (Tables S4-S7).

3 Results

Figure 3 plots modeled versus the measured SSA spectra extracted from singular pixels from a MB, interbed, and veneer (all from site N0/FRT0000BEB8). All spectra exhibit water-ice absorption features at 1.5 μm , 1.5 μm , and 2.0 μm (including the very "dirty" veneer). The spectral fits are very good; the majority of the spectra fit exceptionally well between 1.4–2.5 μm . The modeled spectra struggle to match the spectra between 1.0–1.4 μm . For all pixels across all ROIs at site N0, the mean RMSE are between 0.004–0.012 with

standard deviation of those RMSE means between ± 0.0003 – 0.0038 . At N8 the mean RMSE are between 0.007 – 0.022 with standard deviation of those RMSE means between ± 0.001 – 0.0039 . The endmember abundances, grain sizes, and RMSE for the spectral fit examples plotted in Figure 3 can be found in the figure caption.

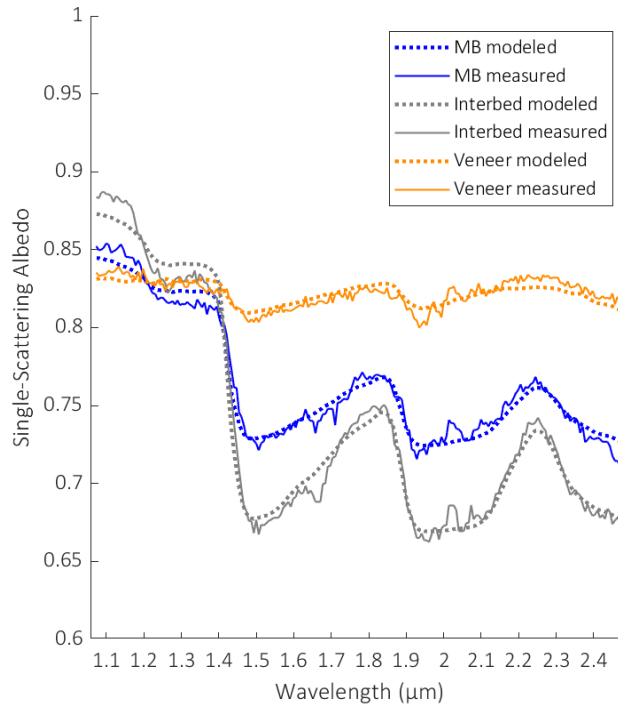


Figure 3. Spectral plot showing examples of measured SSA vs. modeled SSA spectral fits from a 1x1 pixel from a marker bed (MB3), ice-rich interbed pixel, and a veneer pixel (site N0/FRT0000BEB8). MB pixel (x, y) coordinate 483, 785. MB pixel abundance and RMSE: 77% ice, 20% palagonite, 3% dust, RMSE 0.006476. MB pixel grain sizes: 1 mm, 225 μm , 16 μm respectively. Coordinate of interbed pixel: 458, 774. Interbed pixel abundance and RMSE: 87% ice, 9% palagonite, 2% dust, RMSE 0.008198. Interbed pixel grain sizes: 930 μm , 240 μm , 15 μm respectively. Coordinate of veneer pixel: 657, 420. Veneer pixel abundance and RMSE: 35% ice, 60% palagonite, 5% dust, RMSE 0.004337. Veneer pixel grain sizes: 700 μm , 130 μm , 15 μm respectively.

3.1 Intertrough variability

The t-test results for the marker beds at N0 and N8 suggest that all but one of the marker beds have statistically significantly different mean lithic/ice abundances (Table S4, Figure 4). However, while the MBs are compositionally different between the two sites, they have lithic abundances that follow a similar trend stratigraphically down the trough. In other words, at both locations, there is an apparent increase in lithic

abundance as a function of depth for the marker beds. The mean lithic abundance of both sites ranges between 2% and 60% (Figure 4).

At site N0, ice grain sizes across all marker beds have means ranging from ~ 800 μm –1.1 mm, while at N8, the same marker beds have means ranging from ~ 500 μm –800 μm (Table S4, Figure 4). These ice grain size estimates are the first of their kind for specific layers of the NPLD, but surprisingly—and despite the inherent error in Hapke modeling (Li and Li, 2011; Li and Milliken, 2015; Lapotre et al., 2017a, 2017b; Harris and Grindrod, 2018)—they agree with estimates of the ice-rich NPRC (the residual ice cap surface) during late spring/summer from previous studies using different grain size retrieval methods (Langevin et al., 2005; Calvin et al., 2009; Brown et al., 2012). The standard deviation of ice grain sizes is much more significant for marker beds at N0 than at N8. N0 standard deviations in ice grain size range from 140–630 μm from their respective means, while the MBs at N8 have standard deviations less than ~ 200 μm . Similar to the abundance results, the t-tests show ice grain size is also statistically significantly different for each MB between troughs. The grain sizes of the lithic components—palagonite and dust—are occasionally statistically similar (e.g., MB1, MB-1, and MB-2). However, these high p-values for lithic grain size are associated with large standard deviations (except for MB1). In all high p-value cases for t-tests performed on grain size, the standard deviation is greater than the mean grain size (Table S5, Figure 4).

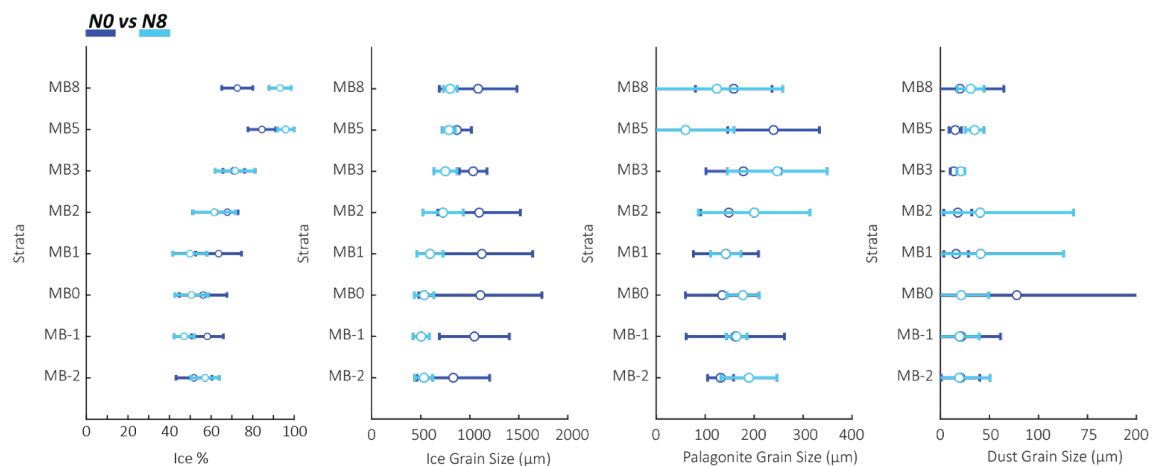


Figure 4. The mean and standard deviation of ice abundance and ice, palagonite, and dust grain size in stratigraphic order. The plots compare the intertrough abundances between sites N0 (dark blue) and N8 (light blue). Lithic abundance is the complementary percentage of ice abundance; therefore, the trend is the opposite with depth.

3.2 Intratrough variability: Dust veneer

T-tests for abundance and ice grain size between the veneered and veneer-free ROIs at site N0 (FRT0000BEB8) have a p-value of 0, meaning the mean compositions of the veneer and the dusty marker bed are statistically significantly different. For the veneer-covered strata, including the interbeds, the lithic abundance is consistent regardless of stratigraphy (means between 63–69%), unlike the veneer-free lithic abundance trend, which ranges from 60–80% and shows a trend with depth (Figure 5). The mean ice grain sizes of the veneered MBs are less than the veneer-free MBs. The only statistically significantly similar ROIs are the dust grain sizes from MB5 and MB2 (Table S7). The veneer and veneer-free ROIs have mean palagonite grain sizes ranging between 100–300 microns (very fine to medium sand-sized). The palagonite grain size estimates for the veneer are more constrained (have smaller standard deviations) than those of the veneer-free MBs and interbeds and have a consistent mean floating around 115 μm .

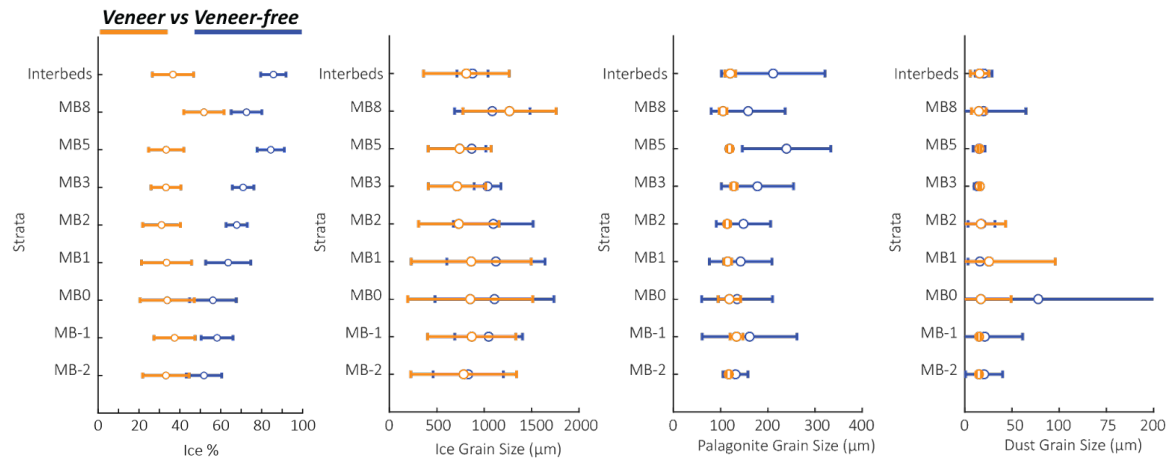


Figure 5. The mean and standard deviation of ice abundance and ice, palagonite, and dust grain size in stratigraphic order. The plots compare the abundances between the veneer-free section of site N0 (dark blue—same values plotted in Figure 4) and the portion of the wall blanketed by a dust veneer (orange). Lithic abundance is the complementary percentage of ice abundance; therefore, the trend is the opposite with depth.

4 Discussion

4.1 Are the MBs expressing inherent layer properties or post-depositional aeolian processes within the trough?

Statistically speaking, all but one of the MBs have distinct abundances between the two sites. The lithic abundance increases as a function of depth/strata at both sites. We believe this supports the hypothesis that surface properties of strata exposed deeper within veneer-free sections of troughs are more heavily affected by superimposed aeolian processes than the upper strata. We suggest the increased lithics with depth reflect transport and re-deposition of sediment (sourced within and outside the trough) with preferential collection towards the bottom of the trough. To further confirm this hypothesis, we pulled abundance profiles down the trough wall (perpendicular to layer strike) for the bright interbed regions from our unmixing results and found the lithic abundance of these regions also increases. Thus, this trend is consistent with depth regardless of layer type, implying it is a process that, in part, depends on the stratigraphic position within the trough.

The one MB that did have statistically similar mean compositions at each site was MB3; this is unsurprising since at both locations, MB3, visually, is the most well-exposed, fresh-looking, and is stratigraphically centered. The trough has the steepest slope in the vertical middle and thus experiences the fastest wind speeds from the katabatic winds flowing perpendicular to the trough wall. Due to these higher wind speeds, the trough middle would be the least likely to hold a lag or surface dust. Any lithics eroded from the upper sections, especially the middle portion experiencing the higher winds, will be transported onto the lower strata as the wind blows it downsection.

With these statistical results in mind, what is the likelihood of looking at inherent layer properties or modern post-depositional/aeolian debris properties? First, we are presented with two possibilities:

- a) the present-day ice-to-lithic ratio of the exposed MB represents the initial ratio when a subsequent NPLD layer capped a MB,
- b) the ice is pore-filling, meaning the lithic sediment either formed at the surface (sublimation lag) or was deposited on the surface (wind-transport) whose pore space is then subsequently filled with ice.

As for the possibility (a), we cannot assume that the compositional properties that were original to deposition of the layer are the same as when the subsequent NPLD layer capped the MB. This is due to the two competing hypotheses of marker bed formation.

1. There was high dust and ice accumulation simultaneously, resulting in the dark-toned layers representing initial depositional composition (Putzig et al., 2009; Laskar et al., 2002).
2. MBs are themselves a build-up of sublimation lag formed when the climate favored deposition of dirty ice and subsequent sublimation annually (Levrard et al., 2007).

However, this implies that layer properties linked to modern post-depositional sublimation lags on the exposed trough wall (i.e., superimposed surface modification) may be spectrally indistinguishable from those linked to paleo post-deposition sublimation lag accumulation (i.e., MB formation mechanism #2). Since the second MB formation mechanism technically does not preserve the true initial layer composition, we refer to surface properties closer to representing the first mechanism as "initial layer properties." Properties more closely resembling sublimation lags we refer to as "post-depositional properties" and do not attempt to distinguish between paleo and modern.

For possibility (b), if the ice percentage reflects post-depositional pore-filling ice, we would expect the ice abundance to be limited to the maximum porosity of regolith or loess-type deposits. A higher porosity would lead to destabilization of the layer exposing it to aeolian erosion. The porosity of the regolith at the Phoenix landing site is estimated to be 50–55% (derived from the average volumetric heat capacity assuming basaltic mineralogy) (Zent et al., 2010). Terrestrial studies of wind-blown loess and paleosols of the Chinese Loess Plateau calculated the porosity to range between 43–57% (paleosol mean porosity of ~48% and loess mean porosity of ~53%) (Zhao et al., 2012 and references therein). These analogs imply that porosity must not exceed 60% if the deposit remains competent and somewhat resilient to being readily transported (unlike loose surface dust). With this information, we pose the following: if a MB has >60% ice (or a lithic abundance between 0–40%), then the high ice percentage cannot be due to pore-filling processes (modern or paleo) as it implies the lag would have been incompetent and removed by wind before the ice could stabilize it. Thus, it is more likely that, for ice-rich MBs, the VNIR derived surface properties are mainly controlled by the initial MB properties linked to simultaneous ice and dust accumulation.

The one MB that is >60% ice and statistically similar between N0 and N8 is MB3. Therefore, we believe the properties reported for MB3 are likely the closest to that of its initial layer properties. MB5 and MB8 also have ice abundance >60%. However, their abundances differ slightly between the two locations. This might be due to a small amount of non-uniform surface debris difference or frost between the sites (although frost is unlikely at this L_s). Surface debris difference is quite apparent upon inspection of the corresponding HiRISE image. We suggest that when determining a MB's initial composition, the consideration of veneer dusting or surface debris must be assessed visually. The results for unmixing in an ice-rich MB can be taken as a minimum for initial ice abundance in a layer since some degree of surface debris is likely. However, due to consistent erosion rates, we propose that an ice-rich layer's inherent properties will be relatively close to this minimum.

4.2 Distinct veneer expression

The brighter interbed regions are more ice-rich than the MBs and likely have inherent lithic abundances closer to that of the bulk NPLD percentage (3%, (Grima et al., 2009)). Therefore, it is believed that these ice-rich beds accumulated during periods of ice stability and low dust deposition (Phillips et al., 2008). In addition, the higher stratigraphy interbeds—which are exceptionally brighter toned and less contaminated by surficial debris than the interbeds deeper within the trough—have model-derived lithic abundances between 2-10%. Thus, these upper strata interbeds are great "control" beds when comparing veneer properties.

Previous studies have suggested that, due to ice absorptions in the spectra of veneer-covered layers, the veneer must be thinner than the skin depth of CRISM and thus the ice absorptions originate from the underlying layer (Calvin et al., 2009). This is corroborated by the stratigraphic brightness variation of the veneered layers in HiRISE data. However, our results suggest otherwise. The veneered interbeds' lithic/ice mean ratio is very similar to the veneered MBs. Meanwhile, we find the veneer-free MBs and veneer-free interbeds to be markedly compositionally different. This implies that the ice absorptions are unique to the veneer and are not due to the substrate beneath it. This finding then further implies that the veneer is fairly homogenous (~35% lithics) and is thicker than the optical skin depth of CRISM (hundreds of microns). As for the spectra, our results show the ice percentage of the underlying layer does not affect the ice/lithic ratio of the veneered pixels; we suggest the ice absorptions are due to pore-filling ice within the veneer, which

also prevents the veneer from being easily stripped. Based on the homogeneity of the veneer's composition, the stratigraphic brightness variation in HiRISE may be due to other controlling factors such as roughness and slope of the underlying layers rather than their visible compositional characteristics. Our findings are consistent with and build upon Rodriguez et al., 2007: the veneer is mainly allochthonous debris cemented by pore-filling ice, likely a mixture of eroded and aeolian transported lithics from other troughs plus aggregates of atmospheric mineral dust that settled annually after the polar vortex dissipated.

Despite the inherent error associated with Hapke mixture modeling and the unknown nature of the lithic components, the veneer grain size estimates are surprisingly realistic for the two lithic endmembers ($\sim 115\text{ }\mu\text{m}$ for palagonite and $\sim 15\text{ }\mu\text{m}$ for atmospheric mineral dust). Dust particles in the atmosphere have been measured to be only a few microns in diameter (Pollack et al., 1995). Once settled on the surface, individual dust grains cohere together to form aggregates, resulting in longer optical paths for propagated light. Dust that has settled on the rovers turns into aggregates forming grains with physical diameters anywhere between $4\text{ }\mu\text{m}$ and $250\text{ }\mu\text{m}$ (Vaughan, 2010). Additionally, even though palagonite is cryptocrystalline, the optical path length of palagonite aggregates would be roughly sand-sized (expected to be within the same dust aggregate diameter range found by the rovers).

4.3 Comparisons between radar- and VNIR-derived abundance estimates

MBs exposed in troughs have yet to be genetically linked to the strong radar reflective subsurface layers observed by radar (Christian et al., 2013). Previously, it has been suggested that a single subsurface reflector could be the result of multiple thin dusty layers within a single resolution cell of SHARAD (i.e., multiple MBs close together) (Campbell and Morgan, 2018). However, recent work by Lalich et al. (2019) presented evidence suggesting that multiple thin layers is not likely the dominant mechanism for the strong radar reflectors and that they are most likely indeed MBs; however, we still have no direct link (one-to-one link) between a single strong radar reflector and a MB observed at the surface. Furthermore, tying the elevation data between the radargram and surface topography brings with it a number of technical challenges as well (Christian et al., 2013). Regardless, the SHARAD estimates are closer to actual initial layer properties (and therefore primary accumulation properties) than to surface-derived properties as modern-day surface processes do not influence the SHARAD estimates.

Therefore, we can qualitatively use SHARAD-derived lithic abundance estimates to assess whether the increase in lithic abundance in our results is due to surficial modern aeolian processes/redistribution of transported dust. Lalich et al. (2019) investigated 10 different locations to derive lithic content estimates from SHARAD data. One location, in particular, is very close to trough sites N0 and N8 (Site H in Lalich et al. 2019). The lithic percentage with depth at site H does not follow any apparent trend with depth, unlike the CRISM-derived abundances; this would be consistent with our prediction that this trend is due to the redistribution of lithics via modern-day katabatic winds. However, Lalich et al. (2019) also discovered that SHARAD-derived lithic abundance of a cap-wide reflector (traceable across all 10 of the location) is inconsistent, ranging between 16–40%. They suggest ice and lithic accumulation during initial layer formation is likely to be highly variable depending on local conditions (Lalich et al., 2019). If local paleo-conditions during layer accumulation have such an effect, then, naturally, variability in current local conditions would affect how aeolian processes within troughs affect the expression of these layers (which agrees with results in Chapter 2 of this dissertation). Therefore, the surface expression of the marker beds is two convolved signals of variability: those in paleo-conditions (also represented in radar data) and those in modern-day conditions (represented solely in visible data).

5 Conclusions

We return to the four main questions we framed at the start of this paper, touch upon additional discoveries, discuss implications for the climatic history the NPLD holds, and make suggestions for future research.

What are the lithic and ice abundance and grain sizes (i.e., surface properties) of the NPLD as measured by VNIR spectroscopic techniques?

- The MBs of the well-exposed NPLD (at sites N0 and N8) range from ~97%–40% ice with a decreasing ice abundance with depth down the trough wall at both site N0 and N8.
- Mean Ice grain diameters range from ~500 μm to 1100 μm . Palagonite grain sizes range between ~50–350 μm (very fine to medium-sized sand). Dust endmember grain sizes range between ~10 μm –75 μm (dust/clay to very fine sand-sized).
- Modeled ice grain sizes for the MBs are similar to grain size estimates from previous studies (700 μm –1000 μm ; Langevin et al., 2005; Calvin et al., 2009; Brown et al., 2012).

- For the veneered NPLD, ice-to-lithic ratios are consistent regardless of the underlying layer (mean lithic abundances between 63-69%).

Do VNIR constrained surface properties of regionally continuous MBs significantly vary between well-exposed NPLD outcrops?

- For the majority of MBs, the abundances and grain sizes are statistically different between both locations.
- For MB-1 and MB-2 the dust endmember grain sizes are statistically significantly similar. Likewise, for MB1 and MB-1 the palagonite grain sizes are statistically significantly similar. However, this was only associated with large standard deviations (except for the palagonite in MB1).
- MB3 was the only MB with statistically significantly similar mean abundances. We proposed this is because MB3 is the most well-exposed, fresh-looking, and stratigraphically centered at the steepest portion of the trough wall, thus experiences the fastest katabatic winds flowing perpendicular to the trough wall.

Based on the regional variability, do VNIR constrained surface properties reflect the intrinsic properties of MBs (i.e., properties that represent original MB accumulation processes)?

- We suggest that MBs with VNIR-derived ice abundance >60% more closely reflect or are dominated by their inherent/initial layer properties and, therefore, the paleo-conditions at the time the layer accumulation.
- Only MBs that are both >60% ice abundance and statistically similar should be considered to represent their paleo-conditions confidently. MB3 falls under this criterion. MBs that are not statistically significantly similar between sites must be assessed for the influence of modern-day post-depositional processes as shown with MB5 and MB8.
- The strata with higher dust content (e.g., the strata lower in section) are less likely to be linked to initial layer properties. This is due to the overprinting of the modern-day surface processes (katabatically-delivered surface dust and/or modern sublimation lag), which convolve and dilute the signal of the inherent layer properties.

How do the VNIR abundance estimates compare to the radar estimates?

- The VNIR abundance estimates have an additional layer of complexity compared to the radar results when attempting to deconvolve a climatic signal. They have variability in both paleo-conditions and modern-day conditions, which can impact surface expression.
- While they cannot be genetically linked, the increasing lithic with depth trend shown in the CRISM data is not explicitly apparent in the SHARAD data. This supports that the increase of lithics with depth is not related to inherent layers processes but instead reflects post-depositional modification of the trough surface. However, future studies should take on the challenge of investigating NPLD spectral signals from CRISM collocated with radar signals and HiRISE elevation (georeferenced HiRISE, CTX, SSA CRISM cubes, and unmixed CRISM cubes from this study are available upon request for those interested in the pursuit).

We determined that dust veneers that uniformly blanket trough walls have a distinct and consistent ice-to-lithic ratio and lithic grain sizes regardless of depth in the trough. Because of this, we believe the veneer is optically thick and adhered with pore-filling ice, which stabilizes the veneer over an annual timescale. Further work will use spectrally-derived surface properties to inversely solve for veneer distribution and thickness change on the surface. This would allow future studies to map and obtain the annual contribution of trough sublimation (or lack thereof) to the total polar cap's volatile exchange with the atmosphere and transport out the polar region. In addition, such work would aid in our efforts to understand polar cap evolution and spiral trough migration (e.g., Smith and Holt, 2010, 2015; Spiga and Smith, 2018; Bramson et al., 2019).

This work lays out the level to which the spectral data can determine inherent layer properties based on the compositional results and visual inspection. While the surface expression is inherently a signal mixture of variability in local paleo-conditions during MB formation and the variability in modern-day post-depositional processes, the two signals can be disentangled through compositional unmixing, rigorous statistical comparisons, and careful consideration of the veneer and transported lithics. Future work looking to integrate radar and hyperspectral data could further constrain the MB formation processes and their insight into the climatic history of the Amazonian.

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Supplementary Material

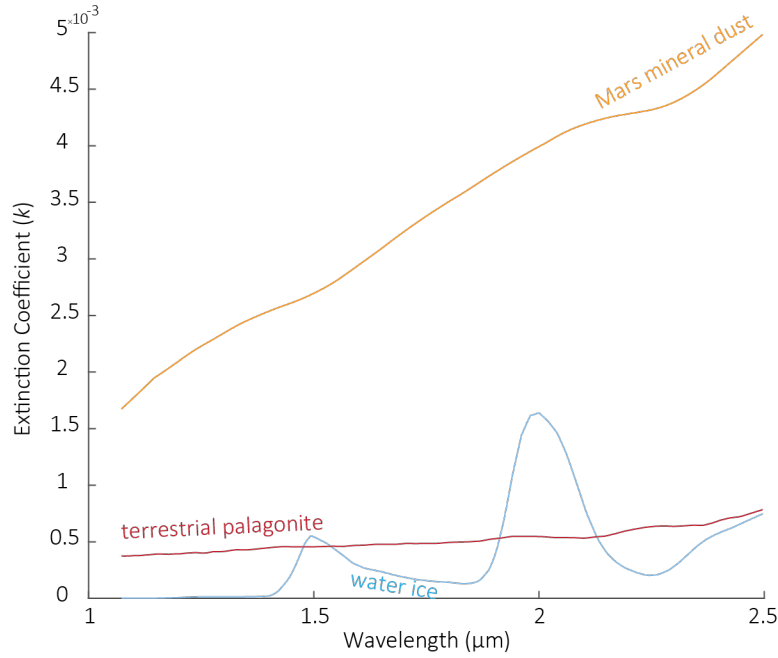


Figure S1. Spectral plot of endmembers optical constant, k (the extinction coefficient), used for the nonlinear spectral unmixing. Endmembers and their source citations are as follows: water-ice (Warren and Brandt, 2008), martian atmospheric dust (Wolff et al., 2006, 2009), and palagonite (Roush, 1993). In the unmixing model the refractive index, n , was assumed to be constant for each endmember. In other words, the average n as a function of wavelength was used for the inverse and forward mixture modeling: $n_{ice} = 1.283$, $n_{palag} = 1.525$, $n_{dust} = 1.488$.

Table S1. weighted means and standard deviations for ice abundance and grain size of the strata at N8 and N0 (including the veneered strata at N0)

Strata	Ice %			Ice grain size (µm)		
	N0 Veneered	N0	N8	N0 Veneered	N0	N8
MB8	52 ±10	73 ±7	73 ±7	1266 ±492	1085 ±397	1085 ±397
MB5	33 ±9	84 ±7	96 ±4	740 ±331	868 ±148	788 ±64
MB3	33 ±7	71 ±5	72 ±10	714 ±303	1035 ±141	750 ±116
MB2	31 ±9	68 ±5	62 ±10	732 ±424	1096 ±420	727 ±207
MB1	33 ±12	64 ±11	50 ±8	863 ±633	1123 ±518	594 ±134
MB0	34 ±13	56 ±11	51 ±8	852 ±659	1078 ±628	534 ±99
MB-1	37 ±10	58 ±8	47 ±5	869 ±465	1047 ±357	504 ±85
MB-2	33 ±11	52 ±9	57 ±7	783 ±558	832 ±370	532 ±91
Interbeds	37 ±10	86 ±6		811 ±452	876 ±164	

Table S2. weighted means and standard deviations for palagonite abundance and grain size of the strata at N8 and N0 (including the veneered strata at N0)

Strata	Palagonite %			Palagonite grain size (µm)		
	N0 Veneered	N0	N8	N0 Veneered	N0	N8
MB8	45 ±9	26 ±8	2 ±4	105 ±8	158 ±78	124 ±134
MB5	63 ±8	12 ±8	1 ±2	119 ±6	240 ±94	60 ±100
MB3	63 ±6	27 ±6	25 ±10	128 ±6	178 ±77	247 ±102
MB2	65 ±9	30 ±5	36 ±11	114 ±6	148 ±58	200 ±114
MB1	63 ±12	34 ±10	47 ±8	115 ±8	142 ±66	142 ±31
MB0	62 ±12	41 ±9	46 ±8	118 ±24	135 ±75	177 ±33
MB-1	58 ±9	39 ±7	50 ±5	134 ±14	161 ±100	164 ±21
MB-2	63 ±10	45 ±8	39 ±7	117 ±6	132 ±26	189 ±57
Interbeds	60 ±10	8 ±7		120 ±11	211 ±110	

Table S3. weighted means and standard deviations for dust abundance and grain size of the strata at N8 and N0 (including the veneered strata at N0)

Strata	Dust %						Dust grain size (µm)					
	N0 Veneered		N0		N8		N0 Veneered		N0		N8	
MB8	2.8	±3.0	1.5	±1.5	4.4	±5.1	15	±8	20	±45	31	±14
MB5	3.6	±2.0	3.8	±6.0	3.6	±3.9	16	±2	15	±6	35	±9
MB3	4.3	±1.4	1.6	±1.0	3.1	±1.6	16	±1	14	±3	21	±4
MB2	3.7	±2.4	1.8	±1.5	2.7	±2.3	17	±26	18	±14	40	±95
MB1	3.5	±2.4	2.1	±1.9	3.1	±2.9	26	±70	16	±13	41	±85
MB0	4.0	±2.5	2.9	±4.2	3.6	±1.8	17	±32	78	±163	21	±28
MB-1	4.2	±1.7	2.5	±1.8	3.3	±1.1	15	±2	21	±40	19	±20
MB-2	3.7	±1.7	3.1	±2.0	3.7	±1.1	15	±3	20	±20	19	±31
Interbeds	3.7	±2.1	5.9	±7.4			16	±10	20	±9		

Table S4. t test results for N0 vs N8: Ice

Strata	Number of pixels		Ice %: weighted mean & std dev		t-value	p-value	Ice grain size (µm): weighted mean & std dev		t-value	p-value
	N0	N8	N0	N8			N0	N8		
MB8	726	442	52 ±10	73 ±7	38	0	1266 ±492	1085 ±397	6.5	0
MB5	475	2267	84 ±7	96 ±4	49.7	0	868 ±148	788 ±64	18.6	0
MB3	849	2775	71 ±5	72 ±10	1.8	0.068	1035 ±141	750 ±116	59.1	0
MB2	320	982	68 ±5	62 ±10	10.4	0	1096 ±420	727 ±207	20.8	0
MB1	293	3433	64 ±11	50 ±8	26.6	0	1123 ±518	594 ±134	44.9	0
MB0	857	5102	56 ±11	51 ±8	17.7	0	1078 ±628	534 ±99	61	0
MB-1	649	2299	58 ±8	47 ±5	44.7	0	1047 ±357	504 ±85	66.5	0
MB-2	704	4493	52 ±9	57 ±7	18.6	0	832 ±370	532 ±91	46	0

Table S5. t test results for N0 vs N8: dust and palagonite grain size (# pixels per ROI same as Table S4)

Strata	Palagonite (µm): weighted mean & std dev		t-value	p-value	Dust (µm): weighted mean & std dev		t-value	p-value
	N0	N8			N0	N8		
MB8	158 ±78	124 ±134	5.3	0	20 ±45	31 ±14	4.2	0
MB5	240 ±94	60 ±100	36.1	0	15 ±6	35 ±9	12.9	0
MB3	178 ±77	247 ±102	18.2	0	14 ±3	21 ±4	10.7	0
MB2	148 ±58	200 ±114	7.8	0	18 ±14	40 ±95	3.3	0.001
MB1	142 ±66	142 ±31	0.2	0.845	16 ±13	41 ±85	4.7	0
MB0	135 ±75	177 ±33	27.4	0	78 ±163	21 ±28	17.6	0
MB-1	161 ±100	164 ±21	1.2	0.247	21 ±40	19 ±20	0.5	0.609
MB-2	132 ±26	189 ±57	26.4	0	20 ±20	19 ±31	0.5	0.598

Table S6. *t* test results for N0 Veneer Covered vs Veneer-free: Ice

Strata	Number of pixels		Ice %: weighted mean & std dev		t-value	p-value	Ice (μm): weighted mean & std dev		t-value	p-value
	Veneer	Veneer-free	Veneer	Veneer-free			Veneer	Veneer-free		
MB8	726	442	52 ±10	73 ±7	38	0	1266 ±492	1085 ±397	6.5	0
MB5	707	475	33 ±9	84 ±7	110	0	740 ±331	868 ±148	7.9	0
MB3	1100	849	33 ±7	71 ±5	128	0	714 ±303	1035 ±141	28.5	0
MB2	934	320	31 ±9	68 ±5	68	0	732 ±424	1096 ±420	13.3	0
MB1	399	293	33 ±12	64 ±11	33	0	863 ±633	1123 ±518	5.8	0
MB0	1529	857	34 ±13	56 ±11	41	0	852 ±659	1078 ±628	9.2	0
MB-1	996	649	37 ±10	58 ±8	45	0	869 ±465	1047 ±357	8.3	0
MB-2	1496	704	33 ±11	52 ±9	39	0	783 ±558	832 ±370	2.1	0
Interbeds	8494	4013	37 ±10	86 ±6	285	0	811 ±452	876 ±164	8.9	0

Table S7. *t* test results for N0 Veneer vs Veneer-free: dust and palagonite grain size (# pixels per ROI same as Table S6)

Strata	Palagonite (μm): weighted mean & std dev		t-value	p-value	Dust (μm): weighted mean & std dev		t-value	p-value
	Veneer	Veneer-free			Veneer	Veneer-free		
MB8	105 ±8	158 ±78	18.2	0	15 ±8	20 ±45	2.9	0.004
MB5	119 ±6	240 ±94	34.2	0	16 ±2	15 ±6	1.8	0.073
MB3	128 ±6	178 ±77	21.6	0	16 ±1	14 ±3	20.6	0
MB2	114 ±6	148 ±58	18	0	17 ±26	18 ±14	0.4	0.681
MB1	115 ±8	142 ±66	8.2	0	26 ±70	16 ±13	2.3	0.02
MB0	118 ±24	135 ±75	7.9	0	17 ±32	78 ±163	14.1	0
MB-1	134 ±14	161 ±100	8.6	0	15 ±2	21 ±40	4.7	0
MB-2	117 ±6	132 ±26	20	0	15 ±3	20 ±20	10.5	0
Interbeds	120 ±11	211 ±110	75.6	0	16 ±10	20 ±9	24.2	0

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Chapter 5

Summary and future suggested research directions

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This dissertation has explored disentangling the complex and coupled signals of past vs. present-day polar processes recorded in the north polar ice cap of Mars using a multi-faceted approach that includes imaging and hyperspectral data for spectral and morphometric analysis and thermal modeling of ice stability. It set out to answer two outstanding questions:

- 1) What are the timescales, spatial scales, and magnitudes of water vapor leaving Mars' north polar spiral troughs, and how have post-depositional surface conditions affected these over the past 300 kyr?
- 2) What are the surface properties of main sequence marker beds of the NPLD and the young, discontinuous dust veneer deposits that mantle the vast majority of equatorial-facing trough walls present-day and do marker bed properties reflect inherent layer properties associated with primary accumulation conditions of the layer or do they reflect post-depositional aeolian activity?

1 Summary of findings and suggested research directions

1.1 The effect of post-depositional processes on the spatio-temporal variability of ice cap evolution

Chapter 2 presented evidence for nonuniform lateral ice loss from equatorial-facing spiral trough walls based on measurements taken from HiRISE topography measurements of marker bed protrusion variability as a function of local trough wall slope and layer packet brightness. From these observations, I showed that a decrease in the amount that a layer (MB) protruded correlated with a decrease in brightness and trough slope. This implies that the surface of equatorial-facing trough walls, which expose NPLD strata, are susceptible to changes in surface conditions during periods of high insolation-driven sublimation periods.

We used a thermal model to calculate sublimation and corresponding lag production rates. I then developed a framework using the model outputs for producing marker bed surface topography that forms positive protrusion off trough walls relative to their recessive icy interbed layers that sandwich them solely from differential sublimation (i.e., not erosionally resistant lag needed). I determined that the key to producing protruding marker beds within this framework is the modulation of lag production versus physical erosion of the lag and the resulting integrated amount of time each stratum spends stable versus sublimating and producing lag. In HiRISE, we measured MBs to protrude between ~1–5 m. Through our calculated cyclical sublimation framework, I found that those protrusions and thus the amount of ice loss experienced

could have developed on the order of thousands of years. At this case study's latitude, the most recent period of time with high enough solar insolation to allow for sublimation of rough walls would have been 60–125 kya. Additionally, I found that the lateral variability in ice retreat that was evident in the HiRISE investigation of this study cannot be the result of spatial variation in solar installation due to trough slope and orientation alone. Instead, lateral variability in ice loss, thus, the present-day protrusion is best explained by the variability in slope wind speed and surface dust removal rate and their combined relationship with the differential sublimation in lag production rate between marker beds and interbeds.

Quite possibly, the most important thing out of Chapter 2's research is the probable importance of having persistent yet transient insulating dust veneers covering entire equatorial-trough wall NPLD exposures during periods of high solar insolation at the north pole when conditions are most favorable for trough wall net ablation. The spatio-temporal distribution and thickness of veneers present-day over seasonal and annual timescales would provide important information to their present-day impact on the surface exchange, or lack thereof, of H₂O. More research on polar dust veneers, particularly explicit mapping of their distribution as a function of time, would allow the first steps needed to more confidently extrapolate the behavior of paleo-veneers' (which are not preserved in the polar record as we know it) contribution to recent trough migration/excavation. In addition, 2D or 3D modeling of trough wall sublimation and erosion (incorporating erosional resistance to MBs) would provide more confident constraints on the polar surface exchange of water vapor and the spatial and temporal scales involved. Continuation of lateral protrusion measurements within the same trough as N0 is also recommended.

1.2 Detangling the expression of inherent layer properties from properties reflecting modern post-depositional aeolian activity

Motivated by (Calvin et al., 2009), Chapter 3 and Chapter 4 sought to use Hapke nonlinear unmixing of CRISM hyperspectral data to approach detangling the expression of inherent properties of regionally correlated NPLD marker beds from properties that reflect post-depositional processes. These chapters lay out the level to which the spectral data can be used to determine inherent layer and dust veneer properties based on the compositional results and visual inspection. Herein we focus on the Chapter 4 findings due to their direct relation to the main thesis question.

The dust veneer had distinct and consistent ice-to-lithic ratio and ice and lithic grain sizes. Furthermore, Hapke spectral unmixing of the dust veneers suggests the veneer is composed of 35% lithics with lithic grain sizes between very fine sand to medium sand-sized. These results confirm Rodriguez et al. (2007)'s hypothesis that veneers are composed of sand-sized particulates held together by pore-ice. Moreover, the abundance in grain sizes for lithic-rich marker beds did not share statistically significantly similar properties to the veneer. This suggests that if marker beds (ones free of dust-veneer deposit) have a protective sublimation lag on their surface, this lag has fundamentally different properties than allochthonous dust veneers that blanket many trough walls.

One main conclusion from this work is that the surface expression is inherently a mixed signal of variability in local paleo-conditions during marker bed formation (e.g., Lalic et al., 2019) and the variability in modern-day post-depositional processes, such as downslope aeolian deposition of dust coatings as well as incomplete blanketing of trough walls by dust veneers. There is a downslope trend of increased lithic abundance at both investigated trough sites. Despite aeolian processes mixing with inherent layer properties for the strata deeper in the trough, I determined at least one marker bed that most likely expresses surface properties closest to their primary accumulation properties (i.e., MB3) at the two trough sites that were investigated.

This work presents the first-ever reported estimates for abundance and grain size of ice and lithics for distinct NPLD strata. More CRISM images covering veneer and veneer-free portions of exposed NPLD co-located with HiRISE (+DTMs) exist but were not included in this study due to project time constraints. Continuation of layer composition and grain sizes quantification of ice and lithic components from these additional co-located data would help further constrain trough and NPLD surface properties. Additionally, I highly encourage large effort collaboration between SHARAD, CRISM, and HiRISE specialists to work together to integrate data products for the best chance to genetically link the subsurface internal radar reflectors to visible stratigraphy at the surface. Other suggested future work could use spectrally-derived surface properties to inversely solve for veneer distribution and thickness change on the surface (e.g., Cull et al., 2010), allowing for future studies to map and obtain the annual contribution of trough sublimation (or lack thereof) to the total polar cap's volatile exchange with the atmosphere and transport out the polar region. Future studies might get us closer to constraining what is recycled back into the cap despite ablating or

eroding off equatorial-facing trough slopes. These types of studies would aid in our efforts to understand polar cap evolution and spiral trough migration (e.g., Smith and Holt, 2010, 2015; Spiga and Smith, 2018; Bramson et al., 2019).

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