

**PLANETARY RADAR AND GEOMORPHOLOGY IN THE
CRYOSPHERE: INVESTIGATING GLACIAL-PERIGLACIAL
LANDSYSTEMS IN PHLEGRA MONTES, MARS AND YUKON
TERRITORY, CANADA**

**RADAR ET GÉOMORPHOLOGIE PLANÉTAIRE DANS LA
CRYOSPHERE: INVESTIGATION DES SYSTEMES GLACIAIRES-
PERIGLACIAIRES DANS PHLEGRA MONTES, MARS ET LE
TERRITOIRE DU YUKON, CANADA**

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Abstract

Mars has abundant water-ice across its surface and in the subsurface, particularly in the form of glacier and permafrost ice. These cryosphere elements drive the development of numerous landforms that are also observed on Earth, enabling comparative studies between both planets. In particular, the mid-latitude region of Mars (30-50°N) is where thousands of *viscous flow features* (VFFs) also called debris-covered glaciers, are situated.

In this PhD thesis, I use a combination of orbital radar sounding, high-resolution imagery, topographic data, and geomorphological techniques to analyze glacial landforms in the *Phlegra Montes* region of Mars. This study is supported by geophysical field investigations in the Canadian Arctic, a well-established analogue for Martian permafrost and glacial landscapes. Together, these approaches provide new insights into the structure, dynamics, and detection of shallow ice systems on Mars and Earth.

Three major findings are presented. First, I report the discovery of a debris-covered glacier with terraced topography as well as a hanging glacier in Phlegra Montes, Mars detected using SHARAD (radar) data, representing the first known observation of such features on Mars. Secondly, I provide evidence supporting past interpretations that VFFs and the *latitude-dependent mantle* (LDM) are separate, yet overlapping systems. Lastly, through field-based radar and sedimentological surveys in Tombstone Territorial Park, I identify the most suitable radar frequency range for detecting shallow subsurface ice (1–5 m), offering direct guidance for future Mars missions using radar instrumentation.

These findings enhance our understanding of Mars' cryosphere and its recent glacial history. They also support mission planning efforts for future exploration and subsurface ice mapping, for example and/or including the *International Mars Ice Mapper (I-MIM)* and *Mars Life Explorer (MLE)* missions.

Résumé

Mars possède d'abondantes glaces d'eau à la surface et dans le sous-terrain, notamment sous forme de glaciers et de pergélisols. Ces éléments de la cryosphère sont à l'origine de nombreuses morphologies qui sont également observées sur Terre, ce qui permet d'effectuer des études comparatives entre les deux planètes. En particulier, la région des latitudes moyennes de Mars (30-50°N) est l'endroit où se trouvent des milliers de caractéristiques de *viscous flow features* (VFF), également appelées glaciers couverts de débris.

Dans cette thèse de doctorat, j'utilise une combinaison de sondages radar orbitaux, d'images à haute résolution, des données topographiques et de techniques géomorphologiques pour analyser les morphologies glaciaires dans la région de *Phlegra Montes* sur Mars. Cette étude est soutenue par des recherches géophysiques sur le terrain dans l'Arctique canadien, un analogue bien établi du pergélisol et des paysages glaciaires martiens. Ensemble, ces approches fournissent des nouvelles informations sur la structure, la dynamique et la détection des systèmes de glace peu profondes sur Mars et sur la Terre.

Trois résultats majeurs sont présentés. Premièrement, je rapporte la découverte d'un glacier couvert de débris avec une topographie en terrasses ainsi qu'un glacier suspendu à *Phlegra Montes*, sur Mars, détecté à l'aide de données SHARAD (radar), ce qui représente la première observation connue de telles caractéristiques sur Mars. Deuxièmement, j'apporte des preuves à l'appui des interprétations antérieures selon lesquelles les VFF et *latitude dependent mantle* (LDM) sont des systèmes distincts, mais qui se chevauchent. Enfin, grâce à des études radar et sédimentologiques menées sur le terrain dans le parc territorial du Tombstone, j'identifie la gamme de fréquences radar la plus appropriée pour détecter la glace de subsurface peu profonde (1-5 m), offrant ainsi une orientation directe pour les futures missions martiennes utilisant des instruments radar.

Ces résultats améliorent notre compréhension de la cryosphère de Mars et de son histoire glaciaire récente. Ils soutiennent également les efforts de planification des missions d'exploration et de cartographie de la glace de subsurface, par exemple et/ou y compris les missions *International Mars Ice Mapper* (I-MIM) et *Mars Life Explorer* (MLE).

Keywords

glacial, periglacial, geomorphology, Phlegra Montes, debris-covered glaciers, radar sounding, synthetic aperture radar (SAR), ground penetrating radar (GPR), Mars, Canadian Arctic, Yukon Territory

Mots Clés

glacial, périglacial, géomorphologie, Phlegra Montes, glaciers couverts de débris, sondage radar, synthetic aperture radar (SAR), ground penetrating radar (GPR), Mars, Arctique canadien, Territoire du Yukon

Summary for Lay Audience

Mars has a lot of frozen water both on and just below its surface, mostly as debris-covered glaciers and permafrost. These icy features shape the landscape in ways that are similar to frozen regions on Earth. One area where this is especially noticeable is in Mars' mid-latitudes (between 30° and 50° north), where thousands of buried glaciers, called *viscous flow features* (VFFs), are found.

In this project, I studied these glaciers in a mountain region on Mars called *Phlegra Montes* using satellite radar, detailed images, and elevation data. I also did fieldwork in the Canadian Arctic, a place on Earth that looks and behaves a lot like Mars' frozen terrain. By combining data from Earth and Mars, I learned more about how to detect and understand shallow ice beneath the surface.

I made three key discoveries. First, I found the first known glacier on Mars with stair-like interior structures and a "hanging" glacier, using radar data. Second, I showed that Mars' buried glaciers and another icy layer called the *latitude-dependent mantle* (LDM) are actually separate features, even though they overlap in some places. Third, I tested radar tools in the Arctic and figured out the best settings for finding shallow ice. This is information that can help in the planning of future missions to Mars.

These results improve our understanding of Mars' frozen regions and help us prepare for upcoming missions that aim to explore the planet and search for water.

Résumé pour Public Profane

Mars contient beaucoup d'eau gelée sur et juste en dessous de la surface, principalement sous forme de glaciers recouverts de débris et de pergélisol. Ces caractéristiques glacées façonnent le paysage de manière similaire aux régions gelées de la Terre. Les latitudes moyennes de Mars (entre 30° et 50° de latitude nord), où l'on trouve des milliers de glaciers enfouis, appelés *viscous flow features* (VFF), sont trouvés.

Dans ce projet, j'ai étudié ces glaciers dans une région montagneuse de Mars appelée *Phlegra Montes* à l'aide de radars satellitaires, d'images détaillées et de données d'élévation. J'ai également travaillé sur le terrain dans l'Arctique canadien, un endroit sur Terre qui ressemble beaucoup au terrain gelé de Mars et qui se comporte de la même manière. En combinant les données de la Terre et de Mars, j'en ai appris davantage sur la façon de détecter et de comprendre la glace peu profonde sous la surface.

J'ai fait trois découvertes essentielles. Premièrement, j'ai trouvé le premier glacier connu sur Mars avec des structures intérieures en forme d'escalier et un glacier « suspendu », en utilisant des données radar. Deuxièmement, j'ai montré que les glaciers enfouis de Mars et une autre couche glacée appelée *latitude-dependent mantle* (LDM) sont en fait des éléments distincts, même s'ils se chevauchent à certains endroits. Troisièmement, j'ai testé des outils radar dans l'Arctique et j'ai trouvé les meilleurs réglages pour trouver de la glace peu profonde. Ces informations peuvent aider à planifier les futures missions sur Mars.

Ces résultats améliorent notre compréhension des régions gelées de Mars et nous aident à préparer les prochaines missions d'exploration de la planète et de recherche d'eau.

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Y’all are truly the *best*.



Figure i. *Bayanihan* (pronounced buy-uh-nee-hun) is a traditional Filipino practice rooted in the word “bayan,” meaning nation, town, or community. It embodies the idea of people coming together in unity and cooperation, working collectively toward a shared goal. Photo credit: <http://mattersmost.deviantart.com>.

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List of Abbreviations

CaSSIS	Colour and Stereo Surface Imaging System
CCD	Charge-coupled device
CCF	Concentric Crater Fill
CCR	Capacitively-Coupled Resistivity
CI	Confinement Index
CSFD	Crater Size-Frequency Distribution
CTX	Context Camera
DEM	Digital Elevation Model
DTM	Digital Terrain Model
EM	Electromagnetic
ERT	Electrical Resistivity Tomography
ESRI	Environmental Systems Research Institute
FSV	Fresh Shallow Valley
GIS	Geographical Information Systems
GLF	Glacier-Like Form
GPR	Ground Penetrating Radar
HiRISE	High Resolution Imaging Science Experiment
HRSC	High Resolution Stereo Camera
IGM	Instructed Glacier Model
I-MIM	International Mars Ice Mapper Mission
IR	Infrared
ISSM	Ice-sheet and Sea-level System Model
LDA	Lobate Debris Apron
LDM	Latitude-Dependent Mantle

LiDAR	Light Detection and Ranging
LVF	Lineated Valley Fill
MARSIS	Mars Advanced Radar for Subsurface and Ionosphere Sounding
MLE	Mars Life Explorer
MOLA	Mars Orbiter Laser Altimeter
MRO	Mars Reconnaissance Orbiter
PAN	Panchromatic
NIR	Near Infrared
RADAR	Radio Detection and Ranging
RTS	Retrogressive Thaw Slump
SAR	Synthetic Aperture Radar
SlimSAR	Scalable, Light, Integrated, Modular Synthetic Aperture Radar
SHARAD	Shallow Radar
TGO	Trace Gas Orbiter
THEMIS	Thermal Emission Imaging System
UAV	Uncrewed Aerial Vehicle
UHF	Ultra High Frequency
VFF	Viscous Flow Feature

Preface

This thesis is dedicated to my Lola Pina, Mama Be, Wayne, Papa, Julian, and Riley.

"Kung may tiyaga may nilaga." – Filipino proverb

CHAPTER 1

1 Introduction

1.1 Overview

Mars is a frozen world where water-ice is abundant at, near, and under the surface. Glacial ice and permafrost ice are prominent features on Earth's and Mars' surfaces. They drive the formation of many landforms that are expressed and studied on both planets (Andersen et al., 2002; Ballantyne, 2018; Carr and Schaber, 1977; Fastook et al., 2008; French, 2017; Whalley and Azizi, 2003). Radar studies performed to date have substantiated the existence of shallow ground ice (Gallagher et al., 2021; Plaut et al., 2009; Séjourné et al., 2019; Stuurman et al., 2016) and have confirmed the existence of debris-covered glaciers on Mars (Berman et al., 2015; Holt et al., 2008; Petersen et al., 2018a).

This PhD research uses a combination of orbital radar and high-resolution imagery to characterize the geomorphology of ice-related landforms. To supplement, a detailed investigation of the relationship between glacial and periglacial environments on Earth, from remote sensing and terrestrial field data of analogue sites in the Canadian Arctic, are a key focus of the project as they offer the most analogous terrains for Mars' (peri)glacial landscapes. A Martian region in particular focus is Phlegra Montes, a mid-latitude mountain range spanning the southern highlands and northern plains. This region has been suggested as a candidate human landing site for Mars exploration. These environments are characterized by extreme seasonal cold temperatures, and either heavy glaciation or hyperarid, polar desert conditions, which will help us to better understand Mars' cryosphere in terms of the current water budget and recent evolutions in its surface and subsurface water cycles. Findings of this research will contribute to proposed future missions such as the International Mars Ice Mapper (I-MIM) and Mars Life Explorer (MLE) Missions.

1.2 Structure of Thesis

This dissertation is divided into eight (8) chapters. Chapter 1 is this *Introduction*, which provides the scope and context to the entire thesis. It briefly outlines the motivations, objectives, and structure of the dissertation. Chapter 2 outlines the methodology, data, and study regions in focus

for the next chapters of the thesis. Chapter 3, entitled *The Cryosphere*, presents a meta-analysis and literature review focused on cryogenic systems on both Earth and Mars, with a particular emphasis on ice-related processes and landforms. These chapters provide the state-of-the-art background information necessary to contextualize and understand the following chapters. These chapters will address the aims and objectives stated in Section 1.2. Chapters 4-6 are formatted in the integrated-article format where the status of each manuscript is outlined in Section 1.4. Chapter 4 specifically investigates debris-covered glaciers on Phlegra Montes, Mars using orbital radar data using the Shallow RADar (SHARAD) instrument, followed by Chapter 5 which focuses on the mapping, inventory, and geomorphological analysis of these same glacial systems in Phlegra Montes, Mars. Chapter 6 integrates ground-based and airborne analogue field data in the Yukon Territory, Canada, including ground penetrating radar (GPR), resistivity, airborne radar, and field measurements to enhance interpretation and detection of subsurface ice using Earth as a technology demonstration and testbed for Mars. Chapter 7: *Synthesis and Future Directions* discusses unresolved issues in Martian cryosphere studies and more open questions raised by my research findings in connection with other literature. This chapter also addresses the limitations of current methods, and opportunities for future exploration, projects, and analysis. Chapter 8: *Conclusion* summarizes the key findings and contributions of this work to planetary science and briefly outlines recommendations for future research.

1.3 Aims & Objectives

The high-level objective of this PhD dissertation is to better understand Mars' current water budget and recent evolutions in its surface and subsurface water cycles, particularly regarding *viscous flow features* (VFFs). The results of this research will have implications on future mission planning and on our understanding of the cryosphere and habitability of Mars. This dissertation consists of three main project objectives:

- **Objective #1:** Investigate regional ice depth and thickness in Phlegra Montes, Mars from radar sounding observations of glacial landsystems using Shallow Radar (SHARAD). Specifically, *how much water-ice is currently stored in the glacial landsystems of Phlegra Montes, and what are the radar-constrained estimates of its depth and protective lag thickness?* This objective and research question targets the quantification of ice

volume and geometry of the glacial base using SHARAD, while testing a new analytical approach to model the overlying debris layer.

- **Sub-objective #1-1:** Conduct a radar survey of all viscous flow features (VFFs) in Phlegra Montes with SHARAD
- **Sub-objective #1-2:** Propose a two-layer analytical model to quantify lag thickness; previous work has only constrained lag thickness based on the SHARAD resolution but not analytically
- **Sub-objective #1-3:** Estimate the total water-ice volume in Phlegra Montes VFFs
- **Objective #2:** Characterize the geomorphology and latitudinal variation of mid-latitude debris-covered glaciers in Phlegra Montes, Mars -- *What geomorphologic and latitudinal factors control the distribution, morphology, and evolution of viscous flow features (VFFs) in Phlegra Montes, and how do these relate to the latitude dependent mantle (LDM)?* This objective aims to address the spatial and morphometric relationships among glaciers, surface processes, and Mars' climatic history.
 - **Sub-objective #2-1:** Create a new and updated regional database and inventory of VFFs in Phlegra Montes
 - **Sub-objective #2-2:** Characterize geomorphologic relationships between physical properties and processes that shape VFFs (i.e. glacier confinement, topographic openness, surface facies association)
 - **Sub-objective #2-3:** Investigate the juxtaposition and possible interaction between the latitude dependent mantle (LDM) and VFFs
- **Objective #3:** Geophysical surveying of ground ice in Tombstone Territorial Park: Implications for the International Mars Ice Mapper (I-MIM) Mission. This objective aims to bridge planetary and terrestrial studies, directly supporting future mission planning in order to address: *How can terrestrial analogue sites, like Tombstone Territorial Park, inform the detection and classification of near-surface ice deposits on Mars, and what geophysical methods are most effective for shallow ice characterization?*
 - **Sub-objective #3-1:** Detect and characterize ice-rich permafrost landforms (i.e. retrogressive thaw slumps, ice-wedge polygons, and buried glacial ice) using integrated remote sensing and field methods.

- **Sub-objective #3-2:** Integrate ground-penetrating radar (GPR), capacitively coupled resistivity (CCR) and airborne radar sounding (e.g. SlimSAR) datasets to detect subsurface ice structures and estimate dielectric properties for differentiating ice types;
- **Sub-objective #3-3:** Determine which GPR frequency(ies) is/are the most suitable for characterizing shallow subsurface ice from 1-5m;
- **Sub-objective #3-4:** Assess permafrost vulnerability and stability for the broader understanding of landscape evolution in Tombstone Territorial Park, Yukon.

1.4 Attribution of Published Work

Parts of this dissertation are adapted from research articles that I have authored or co-authored. This thesis includes material that is either published, under peer review, or based on secondary datasets. Where applicable, full attribution is provided at the beginning of each relevant chapter.

1.4.1 Portions Included from or Based on First-Author Publications

Chapter 4, which focuses on subsurface radar analysis of glacial landforms in Phlegra Montes using SHARAD data, is based on work that I have led as first author. A manuscript based on this chapter is currently in review.

Submitted manuscript citation: Andres C. N. and Smith I. B., (in review). Regional ice depth and thickness in Phlegra Montes, Mars from radar characterization of glacial landsystems using SHARAD. *Journal of Geophysical Research (JGR), Planets*.

Chapter 5, which presents a geomorphological analysis of debris-covered glaciers and VFFs, is also derived from a manuscript currently under peer review. While some figures and methods are shared with the manuscript, this chapter expands on them to include additional interpretation and context.

Submitted manuscript citation: Andres C. N., Conway S. J., and Smith I. B., (in review). Distribution and Surface Morphology of Debris-Covered Glaciers and the Latitude Dependent Mantle (LDM) in Phlegra Montes, Mars. *Icarus*.

Dataset citation: Andres, C. N. (2025). Distribution and Surface Morphology of Debris-Covered Glaciers and the Latitude Dependent Mantle (LDM) in Phlegra Montes, Mars [Data set]. In Distribution and Surface Morphology of Debris-Covered Glaciers and the Latitude Dependent Mantle (LDM) in Phlegra Montes, Mars (Version 1).

Zenodo. <https://doi.org/10.5281/zenodo.15079099>

Chapter 6 is currently in preparation for manuscript submission. Fieldwork in the Yukon and data collection was done by our 2023 field team consisting of me (Chimira Andres), Dr. Isaac Smith, Dr. Denis Lacelle, Dr. Mike Daly, Abi Lee, and Ivan Mishev with the assistance of Adam Gaudreau and in collaboration with Dr. Laura Thomson and Dr. Bernhard Rabus.

In-prep manuscript citation: Andres C.N., Smith I.B., Lacelle D., Daly M.G., Lee A., Mishev I., Rabus B., and Thomson L., (in prep). Geophysical surveying of ground ice in Tombstone Territorial Park, Yukon: Implications for the International Mars Ice Mapper (I-MIM) Mission. Planned submission to: *Permafrost and Periglacial Processes*.

The custom python scripts that I generated to create figures, plots, and process datasets in Chapters 4-6 are made publicly available and can be accessed through GitHub: <https://github.com/chimirandres>. I am the lead author on all chapters and primarily responsible for all research, writing, data collection, digitization, processing, and presentation of results.

1.4.2 Sections Based on My Contributions as a Co-Author

Portions of Chapter 3, specifically the synthesis on Martian mid-latitude glaciation and the evolution of the cryosphere, are included from collaborative work published in Landis et al., 2024, for which I contributed as a co-author. Edits have been made for consistency with the structure and objectives of this thesis.

Manuscript citation: Landis, M.E., Acharya, P.J., Alsaeed, N.R., Andres, C., Becerra, P., Calvin, W.M., Cangi, E.M., Cartwright, S.F.A., Chaffin, M.S., Diniega, S. and Dundas, C.M., 2024. Polar science results from Mars Reconnaissance Orbiter: Multiwavelength, multiyear insights. *Icarus*, 419, p.115794.

Selected portions of Chapter 6, which discusses Earth-based geophysical fieldwork in Tombstone Territorial Park as an analogue for Martian ice mapping, are based on collaborative work submitted for publication of the Mars Polar 2024 conference proceedings (Smith et al., 2025, in review). I contributed to the initial paper structure, fieldwork analogues and mid-latitude glaciation sections. These excerpts are included here with modifications for clarity and integration with the broader goals of the thesis.

Submitted manuscript citation: Smith I.B., Andres C.N., Hibbard S., Nerozzi S., Aguilar R., Buhler P., Cornford B., Tamppari L., and Titus T.N. (in review) Mars as One System: Insights from the 8th International Conference on Mars Polar Science and Exploration. *Icarus*.

1.4.3 Use of Secondary Data and Public Datasets

This thesis also makes extensive use of publicly available datasets from the NASA Planetary Data System (PDS), including SHARAD radargrams, MOLA topography, HiRISE and CTX imagery. Earth-based topographic and geophysical data from the Yukon Geological Survey and Polar Geospatial Center are also used, particularly in Chapter 6. All secondary data sources are cited where used.

CHAPTER 2

2 Study areas and Methodology

2.1 Study sites

2.1.1 Phlegra Montes, Mars

Phlegra Montes is a mountain range located in the northern mid-latitudes of Mars, centered around 165°E, 40°N, northeast of Elysium Mons. Stretching from 30°N to 50°N, this region spans a zone across which near-surface ice stability varies, from stable only on pole-facing slopes at low latitudes to ubiquitous at high latitudes. The Phlegra Montes mountain range runs northward and forms part of the Elysium rise complex, resulting from tectonic uplift during the Early Hesperian to Amazonian period (Fastook et al., 2011; Greeley and Guest, 1987; Hall et al., 1986). The southern edge of Phlegra Montes is positioned near the center of the Elysium rise (28°N, 199°W). In addition, several graben-like structures trend westward and northwestward across the region, with one notable example being a valley system at approximately 40°N that has been filled with debris covered ice (Tanaka, 2005).

Today, the ice found in Phlegra Montes is metastable, a remnant of past climate cycles, which have significantly influenced the region's geomorphology, giving rise to VFFs in the complex topography over time. Phlegra Montes presents an exceptional opportunity to study the cryosphere of Mars across an important latitude range, offering insight into water-ice distribution and its potential for use as an in-situ resource, as proposed by McEwen et al. (2020). Previous research (Dickson et al., 2010; Gallagher et al., 2021; Gallagher and Balme, 2015; Kress et al., 2010) has provided evidence supporting the presence of diverse glacial and periglacial landforms indicative of subsurface ice. These include dissected mantle terrain, interpreted as sublimation-modified ice-rich deposits, and polygonised terrain, which forms through thermal contraction cracking of near-surface ice-rich materials (**Figure 2.1**; Kreslavsky and Head, 2002; Levy et al., 2009). Phlegra Montes, on account of its latitudinal range, elevation, and abundance of ice-rich features, serves as a window into Mars' climatic history, water-ice distribution, and the preservation of subsurface ice.

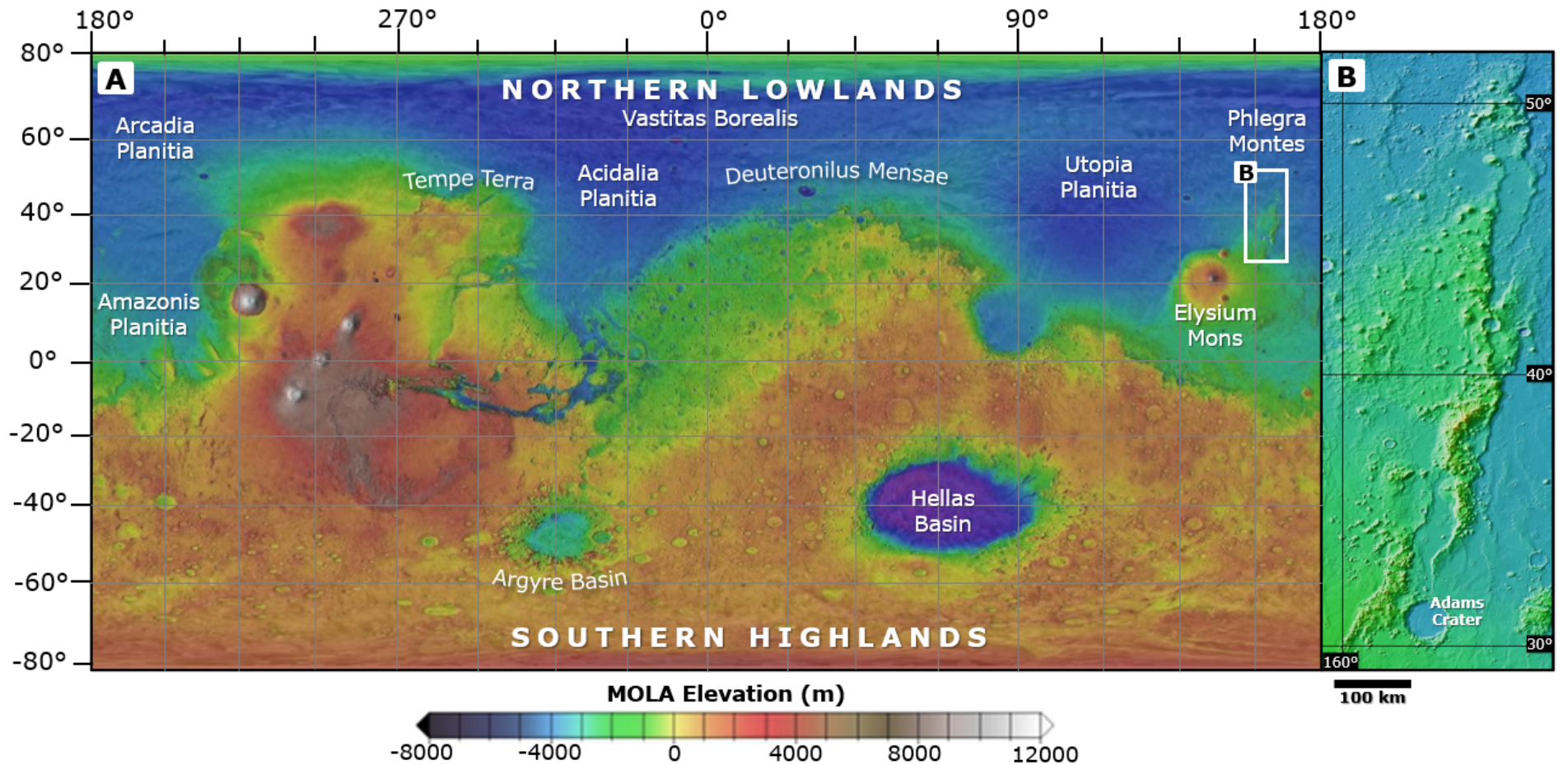


Figure 2.1. (A) A global elevation map of Mars from the Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001), shown in an equirectangular projection overlain with shaded relief. Key regions related to this thesis are labeled. **(B)** Inset map of the Phlegra Montes study area characterized by a ~1500 km mountain range with Utopia Planitia on the west and Amazonis and Arcadia Planitiae on the east.

2.1.2 Tombstone Territorial Park, Yukon Territory

Tombstone Territorial Park is situated in the Ogilvie Mountains, approximately 45 km northeast of Dawson City, Yukon (**Figure 2.2**). The park's indigenous (in Hän) name is *Ddhäl Ch'èl Cha Nän* which means "ragged mountain land." The park is a legacy of the Tr'ondëk Hwëch'in Land Claim Agreement and lies within their Traditional Territory. It spans a range of elevations from 535 to 2,345 meters above sea level (m.a.s.l.), with most of the terrain exceeding 1,000 m.a.s.l. The northern part of the park includes the Blackstone Plateau, composed mainly of Precambrian to Paleozoic metasedimentary and ultramafic rocks (Green, 1972). In contrast, the southern region is underlain by Mesozoic sedimentary and igneous units, including Cretaceous porphyritic syenite that forms the park's highest peaks (Green, 1972). The region contains three principal river systems: the Blackstone, East Blackstone, and Chandindu Rivers, which flow northward and southward, respectively, with their confluence near Chapman Lake. The surficial geology of the park is dominated by thick accumulations of Holocene colluvium (approximately 83%), interspersed with glacial deposits from the Late Pleistocene (around 14%) related to the Reid and McConnell glaciations (Duk-Rodkin and Hinds, 1996). These deposits are expressed as morainal plains and complex till landforms, including hummocky and ridged terrains. Holocene alluvial sediments, comprising coarse sand and gravel with minor finer components, occupy the river valleys (Duk-Rodkin and Hinds, 1996).

Tombstone Territorial Park lies within a continental subarctic climatic zone, typified by short, cool summers and long, cold winters. Meteorological data from a nearby station along the North Klondike River (at 972 m elevation) report a mean annual temperature of -5.4°C (1967–2006), with significant cold periods during winter months when Arctic air masses become trapped in the region (Environment Canada, 2016). Annual precipitation averages around 472 mm, nearly half of which falls as snow. Vegetation is largely composed of alpine tundra, with sparse stands of black spruce and shrub species such as dwarf birch and willow occurring below the treeline and in wetter areas. Permafrost is widespread and generally continuous in the park, particularly in valleys north of North Fork Pass, which marks a hydrological divide between the Arctic and Pacific watersheds. Ground temperature data from nearby boreholes suggest permafrost temperatures between -5 and -2°C . The presence of features such as ice-wedge polygons, thaw slumps, icings, and other cryogenic landforms reflects persistent permafrost conditions since the

Late Pleistocene. Notably, buried glacier ice and relic snow patches dated to over 30,000 years ago have been identified near Chapman Lake and nearby areas. Radiocarbon-dated ice wedges exposed in thaw slumps have provided a detailed chronology of permafrost dynamics, showing transitions between older glacial and more recent Holocene wedges, especially during periods of cooler and wetter climate conditions.

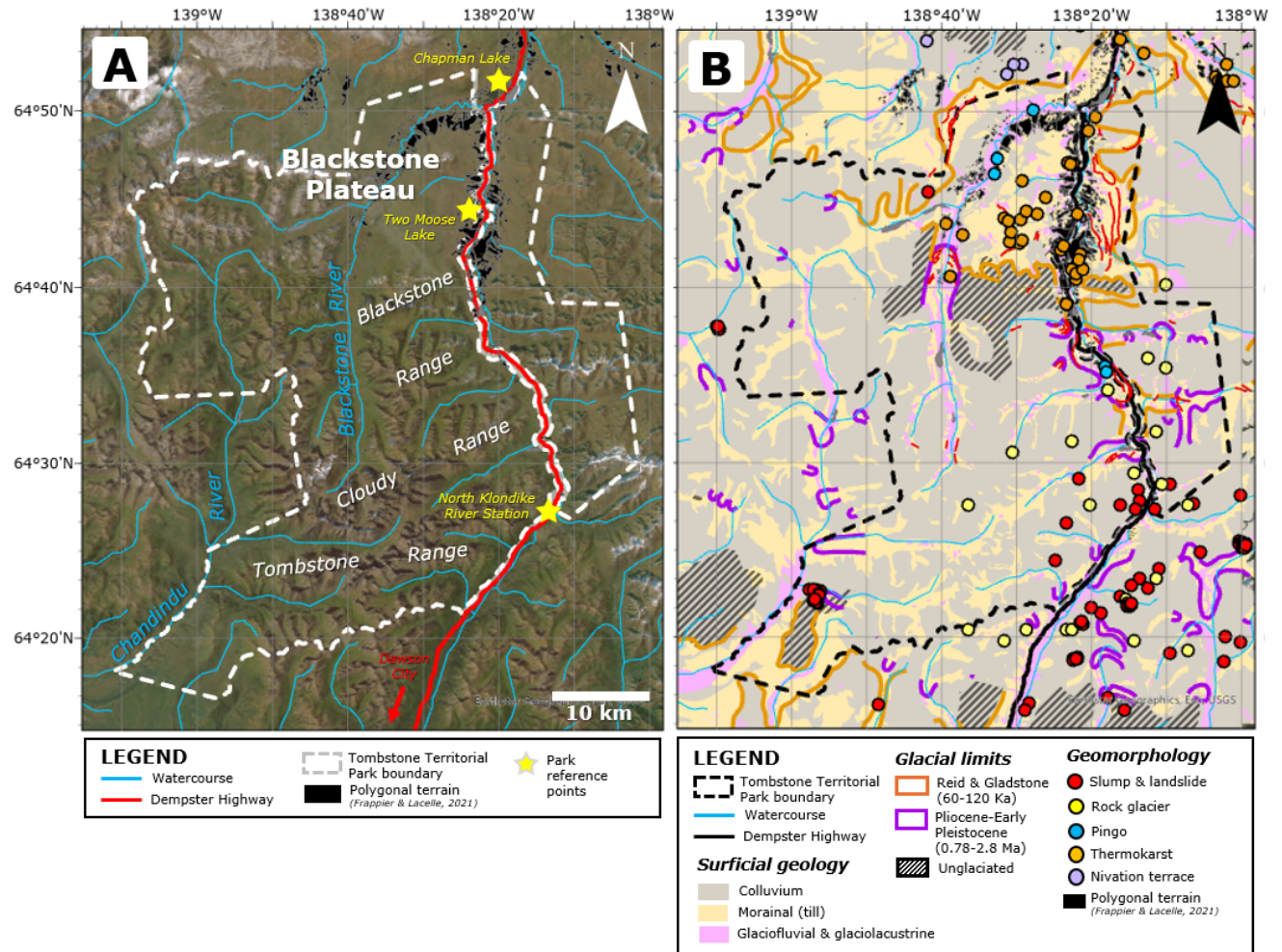


Figure 2.2. Maps showing distribution of ice-wedge polygons in the Tombstone Territorial Park, central Yukon. **(A)** Polygonal terrain and hillshade overlain by semi-transparent MAXAR satellite image (Canadian Digital Elevation Model, Natural Resources Canada, 1945–2011). **(B)** Surficial geology from Yukon Digital Surficial Geology Compilation, Yukon Geological Survey within the Blackstone and East Blackstone River valleys. This map also highlights periglacial landforms taken from Page, 2009; Rampton, 1982. Maps redrawn by Chimira Andres, modified from (Frappier and Lacelle, 2021)

2.2 Methodology

This research integrates a diverse suite of remote sensing datasets and field-based methods to investigate surface and subsurface geological structures across both Mars and terrestrial

analogue environments. The multi-scale approach combines radar sounding, high-resolution imagery, digital elevation models, and ground truthing through field observations. These complementary datasets enable comprehensive geomorphic and subsurface analysis that merges fields of geophysics and geomorphology.

Radar instruments are used for both subsurface sounding and surface imaging. On Mars, the Shallow Radar (SHARAD) instrument aboard the Mars Reconnaissance Orbiter (MRO) operates at a central frequency of 20 MHz and has vertical resolution in free space of ~15 m. The Seisware 10.8 software was used to visualize, map, and analyze SHARAD orbits and radargrams along with CO-SHARPS or the Colorado SHARAD Radar Processing System (more on **Section 2.2.1.1**). MRO operates in a relatively low, near-polar, near-circular orbit around Mars at an altitude of ~ 250-320 km above the surface (Zurek and Smrekar, 2007). On Earth, the airborne SlimSAR system, developed by Simon Fraser University (SFU) and ARTEMIS Inc., operates across ultra-high frequency (UHF), L-, and X-bands, offering adaptable sub-meter to ~10 m resolution from ~800 m altitude. Collaborative fieldwork in April 2023 was done to collect this data. Ground-based GPR surveys also in April 2023, conducted using Sensors & Software Inc. systems, employed at four frequency bands that center on 200 MHz, 250 MHz, 500 MHz, and 1000 MHz (**Table 2.1**), achieving centimeter-scale vertical resolution.

Optical, orbital imagery was used to examine surface morphology and aid radar interpretation. On Earth, basemaps from the ESRI ArcGIS Online database (i.e. MAXAR) was used. Martian data from the Context Camera (CTX; MRO NASA), High Resolution Imaging Science Experiment (HiRISE; MRO NASA), and Colour and Stereo Science Imaging Sensor (CaSSIS; TGO ESA) provided resolutions between 0.3 m and 6 m per pixel (**Table 2.1**), supporting the identification of geomorphic features such as ridges, scarp edges, and potential ice-related landforms.

Topographic data were acquired through both laser altimetry and stereo photogrammetry. On Mars, vertical information from MOLA (Mars Global Surveyor) and HRSC (Mars Express) aided slope and depth analyses. For Earth-based analogues, high-resolution elevation data from the ArcticDEM (Morin et al., 2016) and LiDAR surveys (Yukon Geological Survey) supported detailed surface mapping, radar clutter modeling, and volume estimation of permafrost features.

Geographical information system (GIS) -based mapping and analysis were central to this work. ArcGIS Pro and JMARS was used to visualize, analyze, and digitize geomorphic features from both optical and radar datasets. Python scripts were developed and applied to automate processing tasks, extract topographic profiles, perform spatial statistics, visualize data through plots, and enhance integration across datasets. This geospatial framework enabled consistent data analysis for all studies in this thesis.

Fieldwork was conducted in Tombstone Territorial Park, Yukon Territory, Canada, to carry out a technology demonstration of radar sounding capabilities. Field methods included:

- In-situ measurements, such as depth to permafrost active layer, snow/sediment layer thicknesses, and material composition, to calibrate radar and topographic interpretations.
- Field notes and sketches, recorded during daily surveys, to document relationships between surface morphology, stratigraphy, and environmental context.
- GPR surveys, deployed in selected locations to examine shallow subsurface layering and validate radar penetration limits under varying material conditions.
- Geomorphic mapping and terrain classification, using drone images, visual observations and satellite image basemaps to delineate landforms and surface textures.

These ground-based observations support the remote sensing interpretations and provide essential constraints for analyzing radar-derived properties and subsurface geometry. Together, the integrated methodology leverages the strengths of orbital, airborne, and field-scale tools to interpret complex geological settings with greater accuracy.

Chapters 3, 4, and 5 discuss these methods and techniques in further detail. Chapter 3 focuses on the geophysics of the subsurface with radar sounding using SHARAD, while outlining an analytical method to estimate ice thickness, while Chapter 4 focuses on geomorphology and geospatial analysis using high resolution optical imagery and GIS techniques to understand glacier surface morphology. Chapter 5 combines both geophysical and geomorphological methods in addition to fieldwork to conduct a technology demonstration and experiment on an Earth analogue site for shallow (1-10 m) subsurface ice on Mars.

Table 2.1. Overview of Remote Sensing Platforms, Instruments, and Specifications used in this PhD Research. Symbols are to denote (*) Earth and (⊕) Mars.

Data	Mission	Launch	Instrument	Abbreviation	Band and/or Frequency	Altitude	Resolution (per pixel)
Radar	Mars Reconnaissance Orbiter (MRO) [⊕]	2005	Shallow Radar	SHARAD	15–25 MHz	~250–320 km	Vertical: 15 m (free-space) Horizontal: 300 m
	Simon Fraser University (Canada; ARTEMIS Inc.) [*]	~2010s	Scalable, Light, Integrated, Modular Synthetic Aperture Radar	SlimSAR	UHF, L, X-band	~ 20 km (Airborne)	Vertical: ~1 m-10 m Horizontal: <1 to ~10 m (altitude dependent)
	Sensors & Software Inc. (Canada) [*]	2005	Ground Penetrating Radar	GPR	1000, 500, 250, 200 MHz	Ground-based (~surface)	cm-scale
Imagery	Mars Reconnaissance Orbiter (MRO) [⊕]	2005	Context Camera	CTX	0.5–0.7 μm (Visible/NIR)	~250–320 km	~6 m
	Mars Reconnaissance Orbiter (MRO) [⊕]	2005	High Resolution Imaging Science Experiment	HiRISE	~0.4–1.0 μm (Visible/NIR)	~250–320 km	~0.3 m
	Trace Gas Orbiter (TGO) [⊕]	2016	Colour and Stereo Science Imaging System	CaSSIS	~0.475–0.94 μm (Visible/NIR)	~400 km	~4.5 m
Topography	Mars Global Surveyor (MGS) & Mars Express (<i>Blended dataset</i>) [⊕]	2018	Mars Orbiter Laser Altimeter - High Resolution Stereo Camera	MOLA-HRSC (<i>Blended DEM</i>)	1064 nm (laser) combined with visible stereo imaging	~300-400 km	Vertical: ~10 m Horizontal: 50 m
	Polar Geospatial Center [*]	2016	Arctic Digital Elevation Model	ArcticDEM	Visible stereo satellite imagery	~500 km (Earth orbit)	Vertical: ~0.3-2 m Horizontal: 2 m
	Yukon Geologic Survey [*]	2013-ongoing	Light Detection and Ranging	LiDAR	NIR (typically 905–1550 nm)	~500–3000 m (Airborne)	Vertical: <0.15 m Horizontal: <1 m

2.2.1 Radar

Radar or *Radio Detection and Ranging* is a powerful remote sensing technique that involves transmitting radio waves toward a target and analyzing the returning echoes. Originally developed for military applications during World War II, radar has since become an essential tool in planetary science and Earth observation (Muhleman et al., 1991; Ulaby et al., 1982). Unlike passive sensing methods, which depend on sunlight or thermal emissions, radar is an active technique, providing its own source of illumination. This independence allows radar to function effectively day or night and through clouds, smoke, or planetary atmospheres, making it uniquely suited for studying obscured or shadowed regions on Earth, Venus, Titan, and beyond (Campbell, 2002). In planetary science, radar systems are either ground-based or spaceborne and can operate in monostatic (transmit and receive from the same location) or bistatic (transmit and receive from different locations) configurations. These systems offer unique insights into planetary surfaces and subsurfaces, providing measurements of surface roughness, slope, dielectric properties, and even subsurface layering to depths of meters or kilometers, depending on the wavelength used. The scientific value of radar stems from its sensitivity to the physical and electrical properties of materials, in contrast to optical or infrared sensors which mostly detect surface chemistry (Campbell and Ulrichs, 1969; Elachi et al., 2004; Olhoeft, 1998).

Synthetic Aperture Radar (SAR) measures surface backscatter, making it useful for mapping surface roughness, morphology, and changes over time. Its ability to penetrate clouds and operate day or night makes it ideal for planetary bodies with persistent cloud cover, such as Venus and Titan. SAR is also used for interferometric techniques (InSAR) to detect millimeter- to centimeter-scale changes in surface elevation, though it is not well suited for deriving absolute topography. Examples include Cassini RADAR on Titan and anticipated SAR imaging from missions such as VERITAS and EnVision (Venus). *Radar sounding*, in contrast, transmits lower-frequency signals that can penetrate beneath the surface in dry, low-loss materials such as regolith or ice, revealing buried layers and structures that are invisible in optical imagery. Depth of penetration generally increases with decreasing frequency, though at the cost of coarser spatial resolution. Sounding radars are widely used to investigate subsurface stratigraphy, detect and map ice deposits, and reconstruct geologic history on Earth and other planetary bodies. Both SAR and sounding techniques have profoundly advanced our understanding of planetary geology, climate history, and potential habitability, and will continue to play a central role in exploring planetary surfaces and the near-subsurface (**Table 2.2**).

Table 2.2. Frequency Bands Typically Used in Planetary and Earth Observation Radar (* indicates a future mission).

Band	Frequency Range	Wavelength	Applications	EO Radar Examples	Planetary Radar Examples	Notes
Ka-band	26.5–40 GHz	0.75–1.1 cm	High-res terrain mapping, cloud profiling	CloudSat, UAVSAR (high-freq modes)	-	Above K-band
K-band	18–27 GHz	1.1–1.7 cm	Cloud radar, some weather studies	Experimental airborne sensors	-	From German “kurz”, for short
Ku-band	12–18 GHz	1.7–2.5 cm	Snow depth, glacier studies, ocean altimetry	CryoSat, SWOT (partial)	Cassini RADAR	Under K-band
X-band	8–12.5 GHz	2.4–3.75 cm	Urban mapping, infrastructure, landslides	TerraSAR-X, COSMO-SkyMed	Goldstone, Haystack, VERITAS*	Used in missile guidance e “X” marks the spot
C-band	4–8 GHz	3.75–7.5 cm	Agriculture, land cover, snow mapping	ESA Sentinel-1, RADARSAT-2, TSMM*	Mini-RF	A compromise between X- and S-band
S-band	2–4 GHz	7.5–15 cm	Vegetation mapping, soil, wetland studies	RISAT-1, NovaSAR	Arecibo, Goldstone, Apollo 14/15/16, Venera 15/16, Magellan, Mini-RF, EnVision*	Short wave
Ultra high frequency (UHF) 30 MHz – 3 GHz						
L-band	1–2 GHz	15–30 cm	Vegetation, forest structure, soil moisture	NASA NISAR*, ALOS-2, ROSE-L*,	Millstone-Hill, Pioneer Venus	Long wave
P-band	250–500 MHz	60–120 cm	Subsurface mapping, biomass estimation	ESA BIOMASS mission	Arecibo, Apollo 14/15/16	Previous -- applied to early radar systems
HF-band	3–30 MHz	10–100 m	Deep subsurface sounding	–	Apollo 17, MARSIS, SHARAD, Kaguya Lunar Radar Sounder, JUICE*, Europa Clipper*	High frequency (also known as the <i>decameter band</i>)

Modified from (Neish and Carter, 2014)

2.2.1.1 Orbital Radar Sounding: Shallow Radar (SHARAD)

SHARAD (Shallow Radar) is a subsurface radar sounding instrument aboard NASA's Mars Reconnaissance Orbiter (MRO), designed to probe the upper 1-2 km of the Martian crust. Developed by the Italian Space Agency (ASI), SHARAD operates at a central frequency of 20 MHz with a bandwidth of 10 MHz, providing vertical resolution of ~15 m in free space and a depth penetration of several hundred meters to 2 kilometers, depending on the dielectric properties of the subsurface materials. SHARAD data are typically acquired during local nighttime, between approximately 3:00 AM and 4:00 AM local solar time, when ionospheric interference is reduced. Reduced ionospheric electron density at night lessens dispersive effects, which manifest as frequency-dependent time delays that broaden (or “blur”) the surface echo, potentially obscuring shallow subsurface reflectors (Brookner, 1965; Safaeinili et al., 2003). Nighttime acquisition therefore minimizes this dispersion, improving the fidelity of the surface and near-surface returns.

SHARAD's primary objective is to detect and map buried geological structures, particularly those composed of ice, layered deposits, and potentially liquid water. It builds upon the deeper-penetrating Mars Advanced Radar for Subsurface and Ionosphere Sounding (MARSIS) radar (on Mars Express), offering significantly higher vertical resolution but shallower penetration (Seu et al., 2007). This trade-off allows SHARAD to resolve fine internal layering in the polar ice caps, revealing the history of Martian climate cycles and volatile migration (Smith, 2022). Sounding radar is particularly well suited for deriving dielectric properties of subsurface materials (**Figure 2.3; Equation 1**). When both the radar-detected round-trip time delay Δt and the deposit thickness h (from independent topographic data) are known, the relative permittivity ϵ_r can be estimated as:

$$\epsilon_r = \left(\frac{c\Delta t}{2h} \right)^2 \quad (1)$$

where c is the speed of light in a vacuum. Additionally, the loss tangent, which quantifies how radar power is attenuated with depth, can be derived by analyzing the decay in reflected signal strength between interfaces. These dielectric values can then be compared to laboratory data for terrestrial rocks, helping constrain the composition and density of Martian materials. Notably, SHARAD has detected buried glaciers (Holt et al., 2008; Plaut et al., 2009), lava flows (Bharti

and Smith, 2018; Campbell et al., 2021; Carter et al., 2009; Morgan et al., 2015; Simon et al., 2014), sedimentary deposits (Bharti et al., 2022; Campbell et al., 2008; Mishev et al., 2024), and potential subsurface ice reservoirs in mid-latitude regions (Bramson et al., 2015; Stuurman et al., 2016), findings critical to understanding Mars' geologic and climatic evolution.

SHARAD operates in a monostatic configuration with its 10-m dipole antenna and works in synergy with other MRO instruments like HiRISE and CTX, offering a complementary view of surface and subsurface features (Putzig et al., 2024). Like all radar systems, SHARAD data are subject to *sidelobes* (artifacts of processing when doing a time-domain convolution of the transmitted or reference signal and the returned signal) and *clutter* (reflections from off-nadir topography that arrive at the unfortunate time delay that corresponds with potential subsurface reflectors). SHARAD clutter simulations based on MOLA topography help distinguish “real” subsurface reflectors from artifacts (Choudhary et al., 2016). Additionally, CO-SHARPS or the Colorado SHARAD Radar Processing System allows users to apply custom processing parameters to SHARAD data (Putzig et al., 2016).

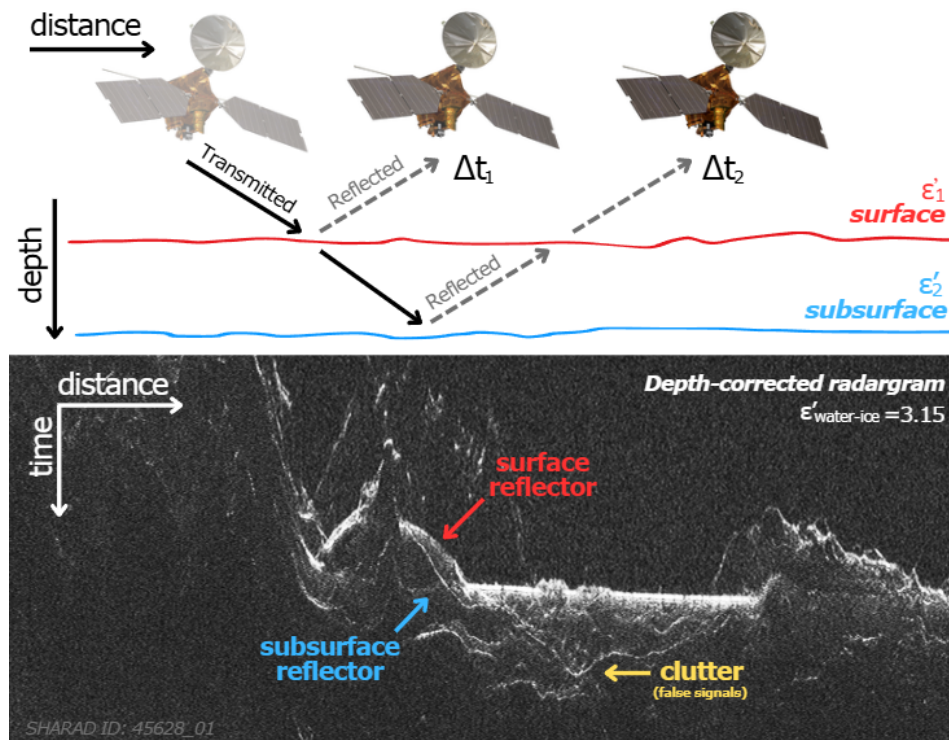


Figure 2.3. (Top) Schematic of SHARAD sounding from orbit as the signal is transmitted and reflected from the surface and subsurface. **(Bottom)** An output radargram of SHARAD. As SHARAD orbits Mars, it emits radar pulses toward the planet. Part of each pulse reflects off the surface, while the rest penetrates the ground. When the signal encounters a boundary between materials with different dielectric properties (ϵ_1 , ϵ_2), it reflects back. By analyzing these reflections, we can quantify certain subsurface material properties. For example, water-ice typically has a dielectric constant of 3.15. Redrawn by Chimira Andres (adapted from The Planetary Society, original illustration by Stuurman, 2020).

2.2.1.2 Synthetic Aperture Radar (SAR)

Synthetic Aperture Radar (SAR) is an active remote sensing technique that transmits microwave pulses toward the surface and records the returning signal after interaction with Earth or planetary terrain. Unlike radar sounding, which produces vertical profiles of the subsurface, SAR generates map-view images of surface backscatter. Operating independently of sunlight, SAR can penetrate clouds, smoke, and even vegetation, providing all-weather, day-and-night imaging capabilities. SAR achieves high spatial resolution by using the motion of the radar platform to synthesize a much larger antenna aperture. This “synthetic aperture” technique (also used in radar sounding) allows airborne or satellite-based systems to capture fine spatial detail without requiring a physically large antenna. Resolution is determined by several factors, including the center frequency, bandwidth, sampling rate, and the processing algorithms applied to the collected data (Curlander and McDonough, 1991; Moreira et al., 2013).

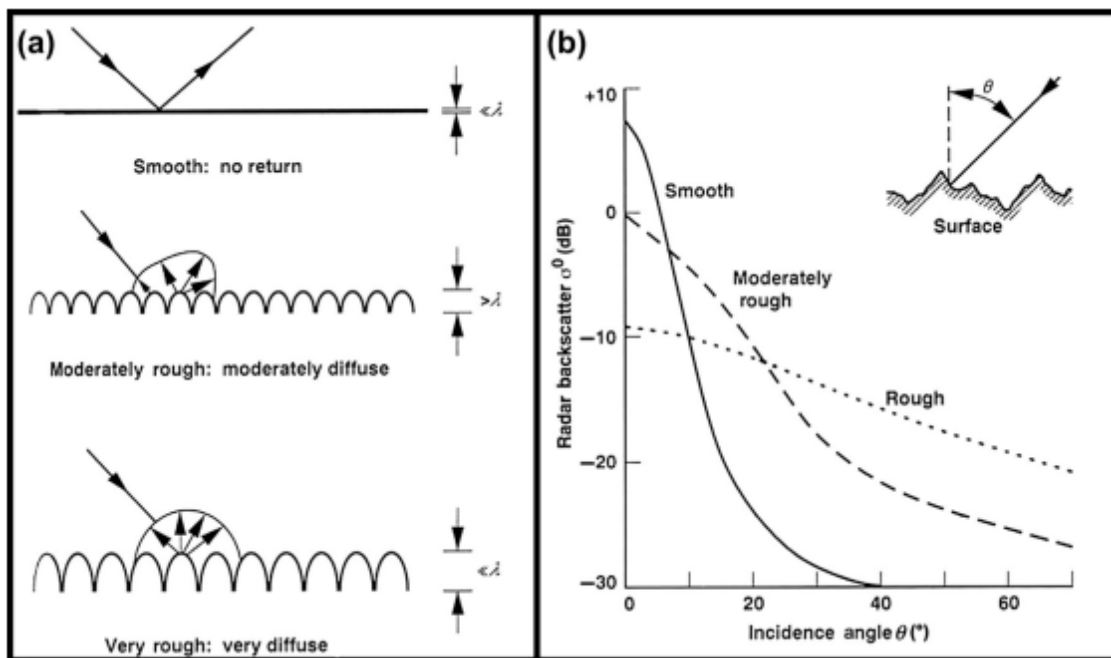


Figure 2.4. (A) The way a planetary surface scatters radar energy depends on how rough it is at the scale of the radar wavelength. Smooth surfaces reflect energy mainly in one direction (specular reflection), often away from the radar receiver, while rough surfaces scatter energy in multiple directions, including back toward the receiver. **(B)** Radar backscatter also changes with the angle at which the radar beam hits the surface. Most geologic materials have some degree of roughness, while smooth surfaces reflect more strongly when the beam is near the specular angle but show little to no return when the incidence angle exceeds about 30°. Figure from (Neish and Carter, 2014).

Although radar images can look similar similar, they carry fundamentally different information. Optical sensors detect reflected sunlight and are sensitive to both chemical composition (via spectral absorption features) and certain physical properties such as slope, aspect, and surface texture, which can influence shading and reflectance. Radar, by contrast, primarily responds to

the electrical (e.g. dielectric constant, conductivity) and physical (e.g. surface roughness, slope) properties of a target. When radar waves hit a material, they excite charges that re-radiate energy based on surface characteristics such as slope, roughness, dielectric constant, and conductivity (Brown, 1967; Curlander and McDonough, 1991; Fitch, 2012; **Figure 2.4**). These factors influence the radar image's albedo and polarimetric response. A surface's radar brightness (i.e. backscatter) is especially sensitive to roughness at the scale of the radar wavelength. Smooth surfaces reflect energy away (specular reflection), while rough surfaces scatter energy in many directions (diffuse scattering), with only a fraction returning to the radar. These scattering behaviors reveal fine details about terrain structure (Fitch, 2012; Franceschetti and Lanari, 2018).

SAR systems can also transmit and receive radar signals in different polarizations such as, horizontal (H) or vertical (V), providing further insight into surface and subsurface properties. The most common configurations include HH (horizontal transmit, horizontal receive), VV (vertical transmit, vertical receive), and cross-polarizations such as HV and VH. Each polarization combination interacts differently with surface features: for instance, HH and VV are often more sensitive to surface roughness and orientation, while HV and VH can highlight structural complexity or volume scattering from vegetation or debris layers (Brown, 1967; Neish and Carter, 2014). Analyzing these polarimetric responses helps to differentiate material types, detect layering, and interpret terrain characteristics more accurately.

One example of a compact and versatile SAR system is SlimSAR, developed by Simon Fraser University and ARTEMIS Inc. SlimSAR is a modular radar that supports operation across multiple frequency bands (e.g., L-, X-, and UHF bands) and is optimized for use on lightweight aircraft or drones (**Table 2.2**). It offers sub-meter horizontal resolution from altitudes ~800 m, making it highly effective for applications such as geological mapping, ice monitoring, and environmental studies (ARTEMIS Inc., 2000). Its low power usage and light payload make it especially suitable for diverse airborne missions. SAR instruments have also been adapted for planetary research on Venus and Titan (one of Saturn's moons).

2.2.1.3 Ground-based: Ground penetrating radar (GPR)

Ground Penetrating Radar (GPR) is a non-invasive geophysical technique used to image the subsurface by transmitting high-frequency electromagnetic (EM) pulses into the ground and

recording the reflected signals. The system comprises a transmitting antenna (Tx) that emits EM pulses into the ground and a receiving antenna (Rx) that detects the signals reflected back from subsurface structures (**Figure 2.5**). When these EM waves encounter interfaces between materials with differing EM properties, such as dielectric permittivity (ϵ), magnetic permeability (μ), and electrical conductivity (σ), a portion of the wave energy is reflected to the surface, while the remainder continues to propagate downward. The analysis of these reflected signals enables the delineation of subsurface structures, including geological formations, stratigraphic layers, voids, and buried objects. The effectiveness of GPR is influenced by several key factors:

1. *Volume Scattering*: Small-scale heterogeneities within the subsurface, such as rocks or voids smaller than one-quarter of the EM wavelength, can cause scattering of the GPR signal. This scattering disperses the wave energy, reducing the coherence of the reflected signals and potentially obscuring the detection of target features.
2. *Attenuation*: As EM waves propagate through the subsurface, their amplitude diminishes due to energy loss mechanisms such as absorption. The attenuation rate is quantified by the attenuation constant (α), which depends on the material's conductivity and permittivity. Higher conductivity materials, like clay-rich soils, exhibit greater attenuation, limiting the depth of GPR penetration.
3. *Skin Depth*: This parameter defines the depth at which the electromagnetic wave's amplitude reduces to $1/e$ ($\approx \frac{1}{2.718} \approx 0.3679$ or approximately 37%) of its original value. Skin depth (δ) is inversely related to the constant attenuation ($\delta = \frac{1}{\alpha}$) and provides an estimate of the GPR's effective penetration depth. Essentially, this is the depth at which the electromagnetics wave's amplitude has decreased to ~37% of its original value due to attenuation.

The pulseEKKO® Pro system by Sensors & Software Inc. exemplifies advanced GPR technology, operating at frequencies ranging from 25 MHz to 1000 MHz (**Table 1.3**). Higher frequencies in this system achieves centimeter-scale vertical resolution, making it particularly effective for detailed imaging of shallow subsurface geology, ice stratigraphy, and engineered structures. Its modular design allows for customization of antennas and software settings to meet specific scientific or engineering requirements.

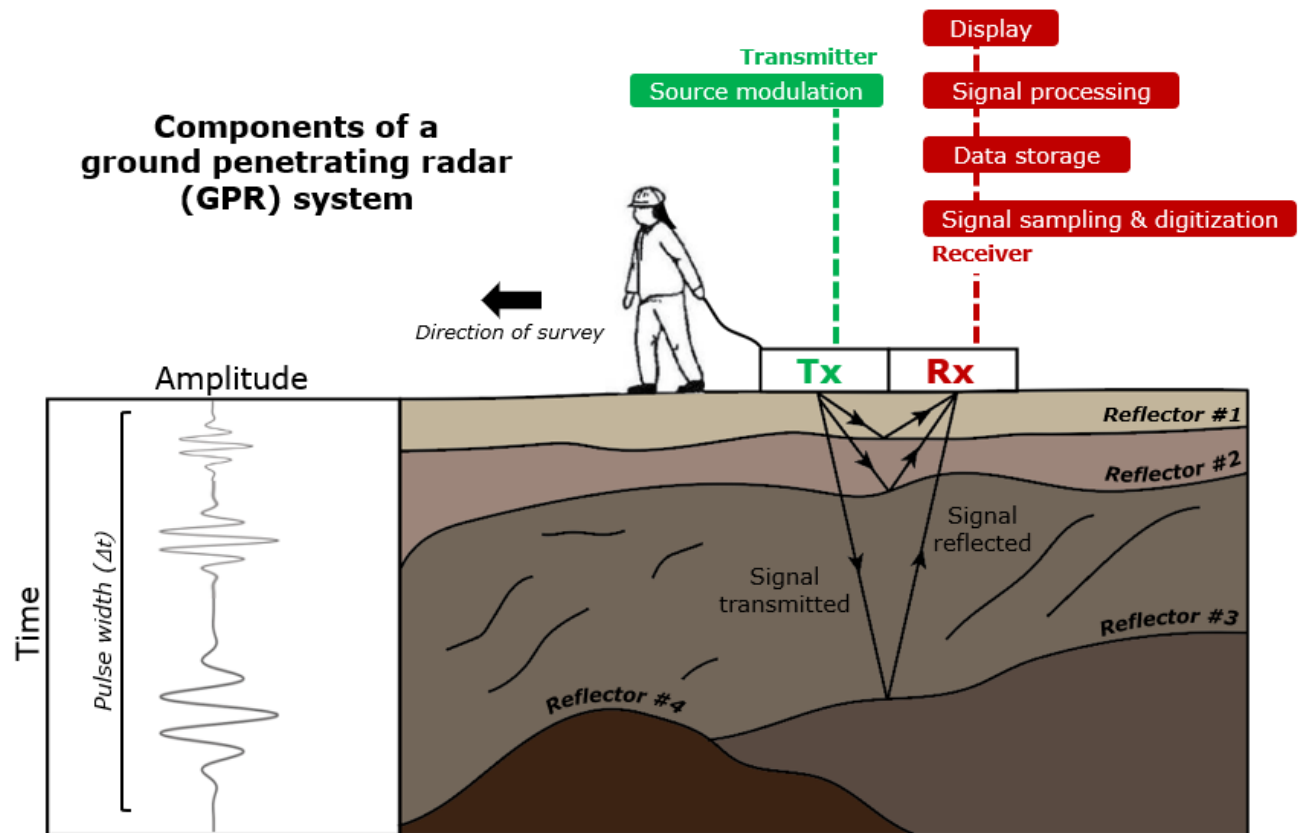


Figure 2.5. Schematic of a common-offset GPR setup. (Right) Various reflector stratigraphy are shown along with the transmitted and reflected signals. (Left) Example radar product of the radar pulse as it travels through material with different dielectric conductivities (y-axis is the two-way time and x-axis is the amplitude). Redrawn by Chimira Andres and adapted from Oldenburg et al., 2017.

Table 2.3. Ground Penetrating Radar (GPR) frequencies and time windows used with their associated depths of penetration, assuming a velocity of 0.1 m/ns (air; adapted from the Sensors & Software pulseEKKO® manual).

Center Frequency (f)	6 dB Bandwidth	Typical Time Window	Wave length ($\lambda = v/f$)	Vertical Resolution ($\lambda/4$)	Approx. Depth of Penetration
50 MHz	25–75 MHz	200–400 ns	2.0 m	~0.6–1.0 m	~30–40 m
100 MHz	50–150 MHz	150–300 ns	1.0 m	~0.3–0.6 m	~15–20 m
200 MHz	100–300 MHz	100 ns	0.5 m	~0.125 m	~10 m
250 MHz	125–375 MHz	~75 ns	0.4 m	~0.07 m	~5 m
400 MHz	200–600 MHz	60 ns	0.25 m	~0.06 m	~3.5 m
500 MHz	250–750 MHz	50 ns	0.20 m	~0.05 m	~2.5 m
800 MHz	400–1200 MHz	~30 ns	0.125 m	~0.03 m	~1.8 m
1000 MHz	500–1500 MHz	25 ns	0.10 m	~0.025 m	~1.25 m

In GPR surveys, the *common offset* method is frequently employed, wherein the transmitting and receiving antennas are maintained at a fixed separation while traversing the survey area (**Figure 2.5**). This approach simplifies data acquisition and is effective for mapping continuous subsurface features. However, it may have limitations in resolving complex structures compared to multi-offset or multi-channel methods. This GPR survey method is discussed in Chapter 6.

2.2.2 Optical Imagery

Optical imagery provides high-resolution observations of surface textures, albedo variations, and landform morphology. In this study, a combination of Mars-based and Earth-based optical datasets is used to identify geomorphic features, monitor surface processes, and support subsurface interpretations from radar and topographic data. **Figure 2.6** illustrates where the major Mars orbital sensors used in this thesis (outlined in Sections 2.2.2.1 to 2.2.2.4) fall along the electromagnetic (EM) spectrum, which is simply a continuous range of energy with oscillating electric and magnetic fields, organized by wavelength or frequency. These sensors span a broad range of wavelengths, from visible, near-infrared (NIR) light to thermal infrared (TIR) and further into the radar portion of the spectrum. Each sensor samples different physical properties of the surface and subsurface, allowing for multi-scale, multi-spectral integration.

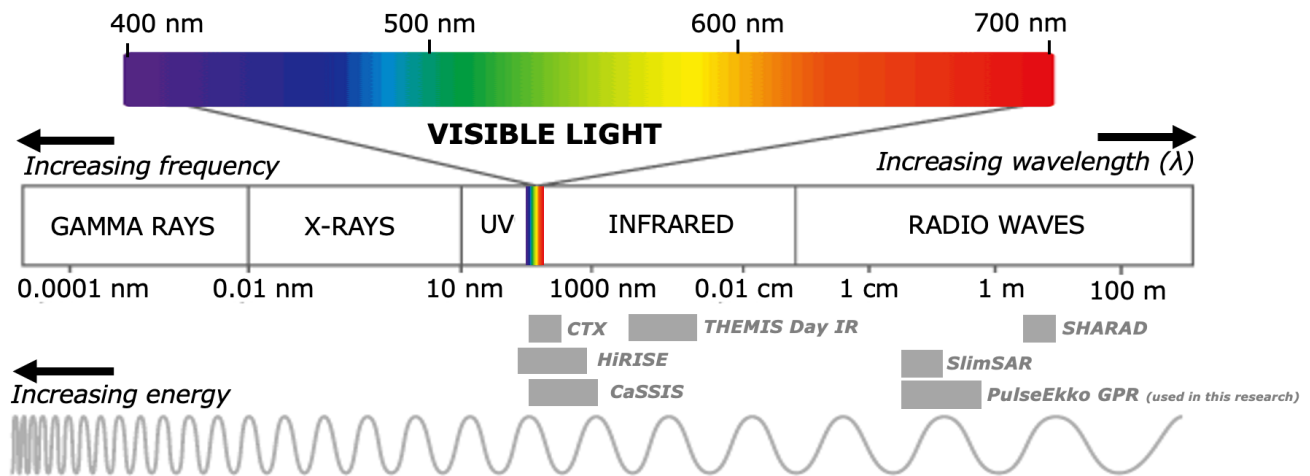


Figure 2.6. Electromagnetic spectrum showing the wavelength ranges of the key remote sensing instruments used in this study. Location of various sensors highlighting their operational wavelength ranges from visible light to radar. See Sections 2.2.2.1–2.2.2.4 for details. Drawn by Chimira Andres.

2.2.2.1 CTX

The Context Camera (CTX) aboard MRO (same spacecraft as SHARAD) captures wide-area, grayscale images of the Martian surface. CTX captures images with a spatial resolution of approximately 6 m/pixel (**Figure 2.7**) over a swath width of about 30 km. It operates in the visible spectrum, specifically between $\sim 0.5\text{--}0.7\ \mu\text{m}$ (**Figure 2.6**), utilizing a 5,064-pixel-wide line array charge-coupled device (CCD) detector (Malin et al., 2007). The camera employs a 350 mm focal length Maksutov-Cassegrain telescope, enabling it to image long strips of the Martian surface as the spacecraft orbits the planet (Malin et al., 2007). CTX (along with all the instruments onboard the MRO) operates from a sun-synchronous orbit, consistently imaging the surface at approximately 3:00

PM local solar time, which ensures high sun angles and consistent illumination for comparative analysis across seasons and years.

In this thesis (Chapters 4-5), the *CTX Global Mosaic (v1)*, developed by Caltech's Murray Lab, served as the primary basemap for surface mapping (Dickson et al., 2023). This mosaic comprises 86,571 individual CTX images acquired between 2006 and 2020, covering over 99.5% of Mars' surface between 88°S and 88°N (Dickson et al., 2023, 2024). The images were registered using a semi-controlled process involving over 3 million control points, ensuring high positional accuracy with absolute errors under 200 m. Dickson et al. (2023) employs a python-based pipeline to preserve the original data integrity, allowing users to trace each mosaic pixel back to its source image. This approach minimizes the risk of misinterpreting blending artifacts as “true” geological features.

The high-resolution and extensive coverage of the CTX Global Mosaic (v1) make it an invaluable tool for identifying geomorphic features, analyzing surface textures, and providing context for radar and topographic data interpretations. Its consistent imaging and broad swath width facilitate the monitoring of surface changes, mapping of large geological structures, and even support the selection of landing sites (Golombek et al., 2016, 2012).

2.2.2.2 HiRISE

The High-Resolution Imaging Science Experiment (HiRISE), also on MRO, is one of the most powerful cameras ever sent to another planet (McEwen et al., 2007). It is a 0.5-m aperture reflecting telescope capable of imaging at resolutions of 0.25 m/pixel (**Figure 2.7**) from an orbital altitude of approximately 300 km. It operates across a spectral range of roughly 0.4–1.0 μm (**Figure 2.6**) and acquires imagery using three distinct channels: BLUE-GREEN (400–600 nm), RED (550–850 nm), and NIR (800–1,000 nm). These bands are recorded by separate CCDs on the instrument's focal plane and are combined into false-colour products that visually enhance subtle variations in surface composition and texture. HiRISE captures a central colour strip with all three channels, which is approximately 1.2 km wide, flanked by broader greyscale coverage from the RED channel, resulting in total image swaths up to 6 km wide. HiRISE has a total of 14 CCDs, RED #10 ceased operation in 2011 and as of early 2025 RED #04 is no longer working, resulting in a vertical black stripe or data gap in some RED images and a reduction of 50% in the colour swath.

HiRISE uses a pushbroom scanning design and has an effective focal length of 12 m, enabling it to acquire long image strips with minimal distortion and high geometric fidelity. Since HiRISE is

supported by precise Spacecraft Planet Instrument C-matrix Events (SPICE) kernel registration and photometric calibration, it is well-suited for identifying stratigraphic relationships, detecting fine-scale layering, and monitoring surface activity.

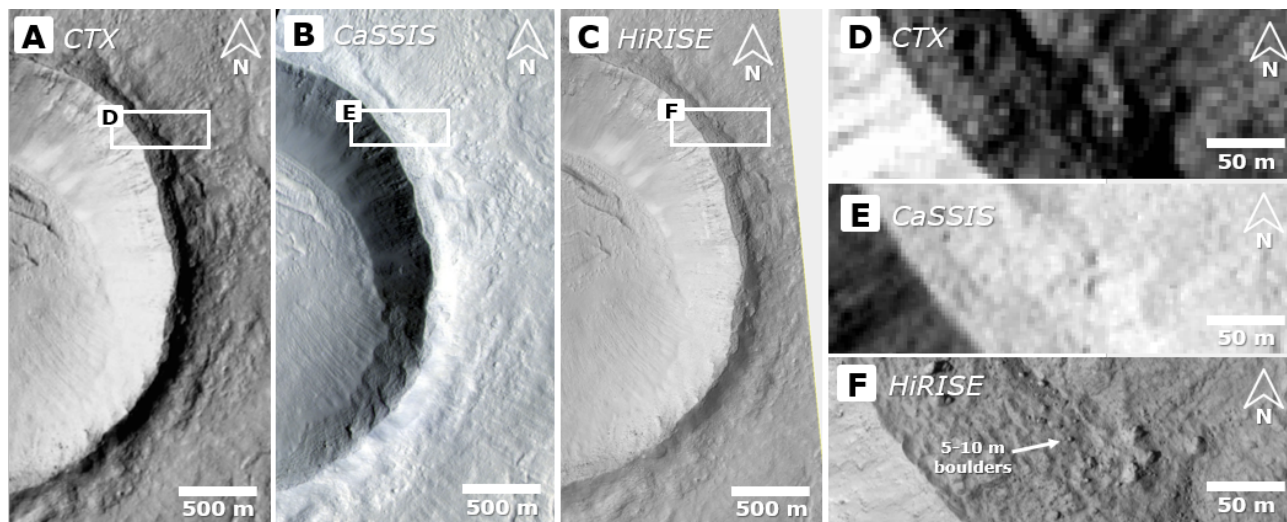


Figure 2.7. A comparison between CTX (6m/px), CaSSIS (4m/px), and HiRISE (0.25m/px). An image of a crater at, 36.8° N, 158.3° E in the eastern part of Phlegra Montes at different resolutions highlighting boulders that can only be seen in HiRISE. (CTX ID G22_026989_2176; CaSSIS ID MY37_024556_039_0, HiRISE ID ESP_043670_2170).

In this thesis (Chapter 5 specifically), HiRISE was used to confirm surface texture classifications (e.g., distinguishing between different types of terrains that are 1-10 m in scale), refine geomorphic boundaries, and provide context for interpreting radar reflectors detected by SHARAD. Its sub meter-scale resolution made it particularly valuable for validating the expression of subsurface features and identifying structures not visible in other optical imagery data. HiRISE coverage was prioritized in areas of geophysical interest; in locations lacking HiRISE data, CTX or CaSSIS imagery was used as the primary reference outlined in Sections 2.2.2.1 and 2.2.2.3 respectively.

2.2.2.3 CaSSIS

The Colour and Stereo Surface Imaging System (CaSSIS) is an instrument aboard the European Space Agency's ExoMars Trace Gas Orbiter (TGO). CaSSIS captures high-resolution push-frame images in both colour and stereo modes, with a nominal spatial resolution of ~4.5 m/pixel at NADIR (**Figure 2.7**). Unlike MRO, TGO is not in a sun-synchronous orbit, so CaSSIS images the surface at varied local times (Thomas et al., 2017). This variability allows CaSSIS to capture features under a range of lighting conditions, which is especially useful for detecting subtle roughness, shadowing, or transient surface activity.

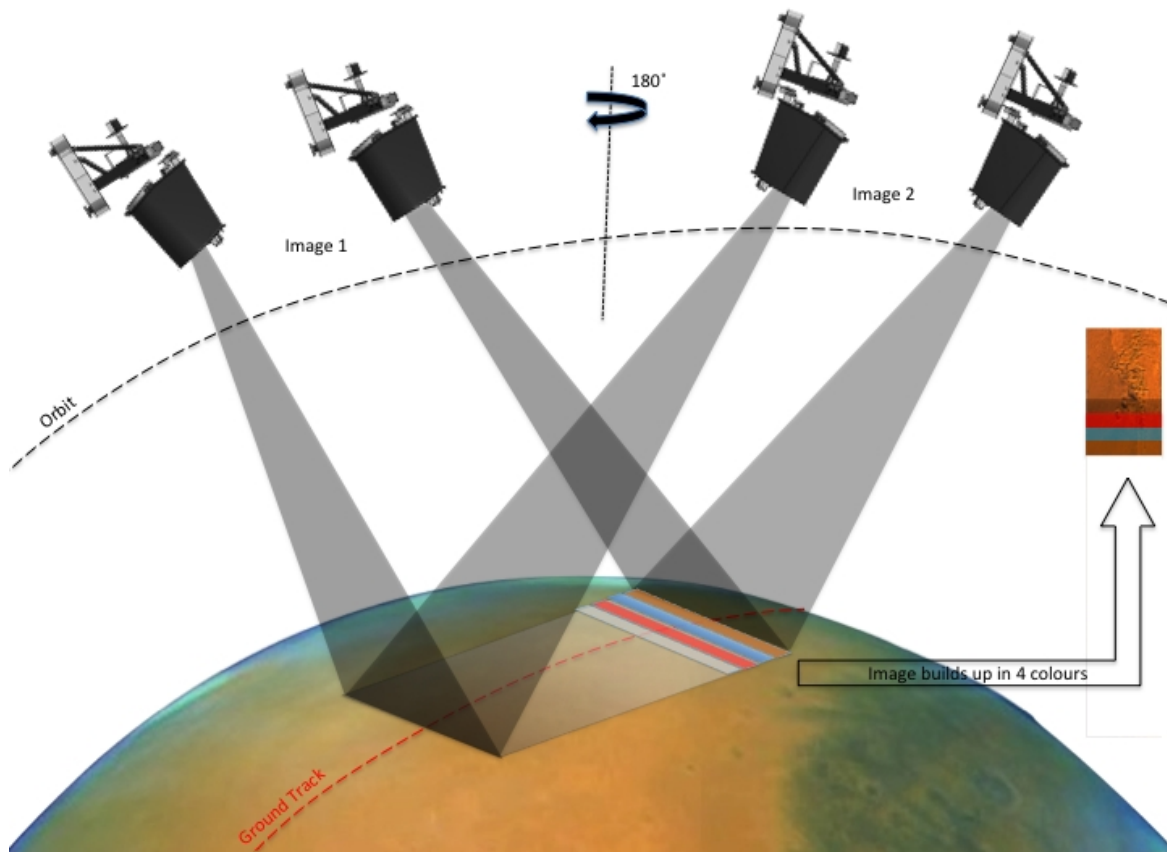


Figure 2.8. Stereo imaging by CaSSIS aboard ESA’s ExoMars TGO. The instrument rotates $\sim 180^\circ$ to acquire stereo pairs, first imaging 10° ahead, then 10° behind the spacecraft. This pivot mechanism, unique among Mars orbiters, enables CaSSIS to generate colour stereo images and DTMs with vertical resolutions of $\sim 10\text{--}20\text{ m}$ (Image from the CaSSIS Instrument Website; Portal Universität Bern).

CaSSIS operates using four broad-band filters: PAN ($\sim 600\text{--}800\text{ nm}$ centered at 675 nm), BLUE (475 nm), NIR (835 nm), and IR (950 nm ; **Figure 2.6**). Colour images are constructed from sequentially captured bands using a push-frame acquisition method (Thomas et al., 2017). The instrument is mounted on a rotation mechanism that enables stereo image acquisition by swiveling the telescope $\sim 180^\circ$ during an orbital pass. This allows CaSSIS to capture one image 10° ahead of the spacecraft’s motion and a second image 10° behind, generating stereo pairs for digital terrain model (DTM) construction (**Figure 2.8**). This design makes CaSSIS the first Mars orbiter camera to incorporate a mechanical pivot for stereo imaging, enabling vertical resolutions of $\sim 10\text{--}20\text{ m}$ depending on lighting and viewing geometry. For this dissertation, CaSSIS was used to supplement CTX and HiRISE data in areas of partial coverage, with its colour imaging supporting the interpretation of surface textures.

2.2.2.4 THEMIS Day Infrared (IR)

The Thermal Emission Imaging System (THEMIS), aboard NASA’s 2001 Mars Odyssey, collects both visible and thermal infrared data. THEMIS operates in a sun-synchronous orbit and used to

acquire imagery at approximately 4:00 PM (daytime IR) and 3:00 AM (nighttime IR) local solar time. However, more recently, its local acquisition time is variable to preserve fuel, while still acquiring observations under varying thermal conditions. The daytime IR data, used in this dissertation (Chapter 4), offers a spatial resolution of ~ 100 m/pixel and spans nine thermal bands between ~ 6.78 and 14.88 μm (**Figure 2.6**). These bands are used to estimate surface thermal inertia, grain size, and albedo, helping distinguish between dust, bedrock, and ice-rich materials. In this dissertation, I only used THEMIS Day IR as a general background layer for visualization and interpretation, particularly in the JMARS software. THEMIS Day IR is also the basemap used in some SHARAD products to show the track overlay.

2.2.3 Topography

Topographic data are essential for analyzing surface morphology, slope, elevation profiles, and the spatial context of radar returns. This dissertation uses Martian and terrestrial elevation models from laser altimetry, stereo photogrammetry, and LiDAR.

2.2.3.1 MOLA/HRSC

Topographic data of Mars used in this study comes from a blended digital elevation model (DEM) product combining measurements from the Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001) aboard NASA's Mars Global Surveyor (MGS) and the High-Resolution Stereo Camera (HRSC; Jaumann et al., 2007) aboard ESA's Mars Express (MEX). This version 2 product was developed to support thermal modeling and surface analyses on Mars (Ferguson et al., 2018) and has a spatial resolution of 200 m per pixel.

MOLA acquired more than 600 million elevation measurements between 1999 and 2001 by firing infrared laser pulses toward the Martian surface and timing their return (Smith et al., 2001). HRSC was the first dedicated stereo camera orbiting Mars (prior to CaSSIS) and consists of 9 CCD line sensors configured to acquire stereo, multicolour, and multi-phase images. To generate DEMs, stereo imaging relies on the principle of parallax, which is the apparent shift in position of surface features when viewed from different angles (Schenk, 2005). HRSC accomplishes this using its 9 CCD line sensors, which are aligned to capture images from slightly different perspectives as the spacecraft moves forward. Its push-broom configuration also allows the instrument to collect stereo pairs (forward, NADIR, and backward views) in a single pass, reducing inconsistencies due to changing illumination. This method is also known as stereo photogrammetry. It identifies conjugate points across the

overlapping images and triangulates their elevation based on the parallax displacement (Jaumann et al., 2007; Schenk, 2005). The resulting DEMs are co-registered and corrected using MOLA altimetry, which provides an absolute elevation reference. Corrections include orbital alignment, pointing refinements, removal of residual distortions, and filtering to smooth artifacts from stereo mismatches. HRSC-derived DEMs can reach grid sizes as fine as 50 m, with vertical accuracy up to 10 m, making them valuable for generating high-resolution DEMs for analyzing topographic trends and landform geomorphology. To produce the blended global product used in this thesis, HRSC stereo-derived elevations were co-registered to MOLA data, which served as a vertical reference frame and was also used to fill coverage gaps. MOLA provided near-global coverage, while HRSC contributes about 40% of the surface of Mars in the v2 blended mosaic (**Table 2.1**; Fergason et al., 2018). The final product uses the high vertical precision of MOLA with the finer spatial resolution of HRSC to create the 200 m/pixel global DEM (Fergason et al., 2018).

2.2.3.2 LiDAR

Light Detection and Ranging (LiDAR) is a remote sensing method primarily used for generating precise topographic models of Earth's surface. Airborne LiDAR systems emit laser pulses and measure the reflected signals to determine surface elevation with high accuracy. The "time of flight" is converted to distance using the speed of light ($c = 3.0 \times 10^8$ m/s), generating a 3D point cloud of elevation data. These irregularly spaced points are then classified to separate ground from non-ground returns (i.e. vegetation and buildings), and ground points are interpolated into a continuous raster to produce DEM. Vertical resolution is typically better than 15 cm, while horizontal resolution can be sub-meter, depending on flight altitude and system configuration.

In this research (Chapter 6), high-resolution LiDAR data collected by the Yukon Geological Survey along the Dempster Highway corridor in 2013–2014 was used to support geophysical fieldwork, GIS mapping, and topographic analysis. This dataset, available through the Government of Yukon's GeoYukon platform, provides sub-meter spatial resolution across a 600–2000 m-wide swath, offering the detail necessary for identifying subtle surface features and informing survey design (Government of Yukon Lidar collection, 2014).

CHAPTER 3

3 The Cryosphere

This chapter provides an overview of the current understanding of the cryosphere on both Earth (Section 3.1) and Mars (Section 3.2). Section 3.1 *Ice on Earth*, is organized into three themes: formation of glaciers, mechanisms of glacial movement, and landforms produced by glacial activity. This section provides the necessary terrestrial background to understand planetary applications on Mars. Section 3.2 *Ice on Mars*, follows the same structure, describing the distribution of ice on Mars, processes that preserve it, and landforms associated with Martian ice with particular focus on viscous flow features (VFFs) or debris-covered glaciers. Finally, Section 3.3 highlights *Open Questions* and gaps in the current knowledge, which motivate the research objectives addressed in this thesis from Section 1.3. I use this chapter to focus on background context information to help in the understanding of the following Chapters 4, 5, and 6.

Comparative planetology enables the application of terrestrial cryospheric knowledge to the Martian environment, where direct observations are more limited. Many glacial and periglacial landforms observed on Earth serve as valuable analogues for interpreting similar features on Mars. Moreover, the techniques used to study ice, such as remote sensing, radar sounding, geomorphological mapping, and geophysical surveys, are often shared across both planetary contexts. These methodological parallels allow for cross-validation of interpretations and refinement of models.

Studying Earth's dynamic cryosphere helps constrain the behavior, composition, and climatic sensitivity of Mars' cryosphere. Mars offers insight into cryogenic processes under low pressure, reduced gravity, and hyper-arid conditions. Section 3.2 *Ice on Mars* will further explore the occurrence, evolution, and detection of ice on Mars, emphasizing how Earth-based analogues and analytical tools inform our understanding of the Martian cryosphere and vice-versa.

3.1 Ice on Earth

3.1.1 A Brief History of Glaciation on Earth

Glaciation has played a central role in shaping Earth's surface, climate history, and the development of life as we know it (Benn and Evans, 2014; Eyles, 1993). The earliest known glaciations date back over 2 billion years during the Paleoproterozoic Era (e.g., Huronian

glaciation), likely linked to the Great Oxygenation Event, which cooled the planet by reducing atmospheric methane. More globally extensive glaciations followed in the Cryogenian Period (~720–635 Ma), with the Sturtian and Marinoan events often referred to as "Snowball Earth" episodes, during which ice may have extended to equatorial latitudes.

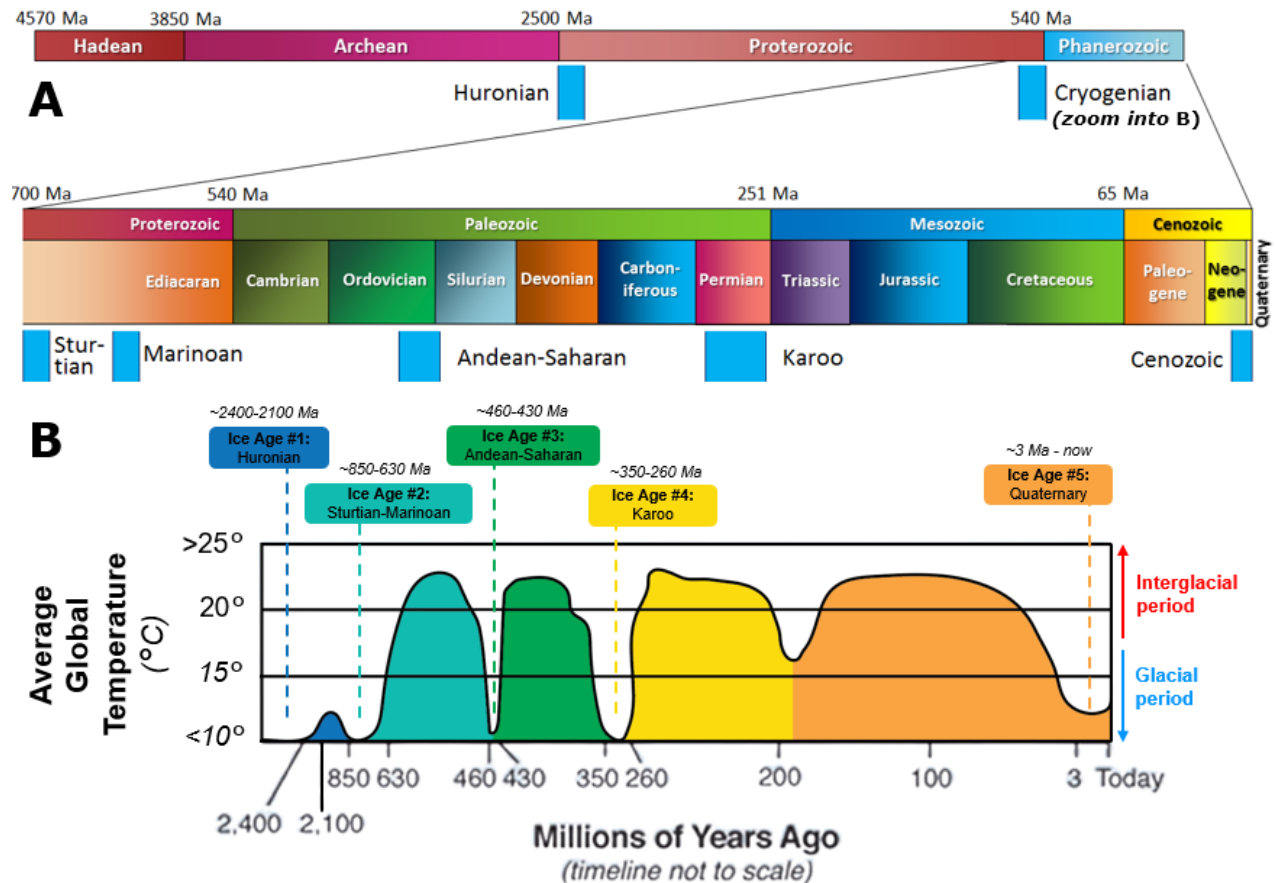


Figure 3.1. (A) Geologic timeline of Earth with the light blue boxes highlighting the glacial periods. **(B)** Zoom into the Phanerozoic, focusing on the five major ice ages that Earth has experienced. We are currently in the middle of a glacial period.

During the Phanerozoic Eon, other significant glaciations include the Andean-Saharan (~450–420 Ma), the Late Paleozoic Karoo glaciation (~360–260 Ma), and the ongoing Late Cenozoic or Quaternary glaciation (**Figure 3.1**) that began approximately 34 million years ago with the formation of the Antarctic Ice Sheet. Each glacial period is marked by complex interactions between tectonic, atmospheric, and orbital forces (Meyers and Malinverno, 2018; Short et al., 1991). The Quaternary Period (>2.6 Ma) is especially important in shaping current landscapes. It is characterized by cyclical glacial and interglacial periods, driven by Milankovitch cycles, long-term changes in Earth's orbit, axial tilt, and precession (Short et al., 1991). These cycles regulate the distribution of solar energy, causing repeated advances and retreats of large ice sheets,

particularly over North America, Europe, and Asia. Similarly, orbital cycles are also important for modulating Mars' cryosphere, see Section 3.2.2.

The Last Glacial Maximum (LGM), around 20,000 years ago, was the peak of the most recent major glaciation (Clark et al., 2009). Since then, Earth has entered the Holocene, a relatively warm interglacial period that began approximately 11,700 years ago. This period has seen a distinct retreat of continental glaciers and formation of many modern coastlines. While current climate trends are influenced by natural cycles, there are several anthropogenic factors (i.e. greenhouse gas emissions) which may be altering the progression toward the next glacial period. The geological record, including glaciomarine and glacio-terrestrial deposits, continues to be the primary data source for reconstructing Earth's glacial history (Benn and Evans, 2014). From 20th-century developments in paleoclimate modeling to isotope and glacial reconstructions, our understanding of glacial cycles is constantly evolving.

3.1.2 Glacial domain: *How do glaciers form and flow?*

Continuous snow accumulation over decades starts the metamorphism from snow to glacier ice. Glaciers form through the gradual compaction of snow into firn, then into ice in regions where snowfall (accumulation) exceeds melting (ablation) rate (Benn and Evans, 2014; Paterson, 1994). In the *accumulation zone*, snow builds up year after year. With time and pressure, the snow compacts into firn, a dense, granular form, and eventually into solid glacial ice due to the increasing pressure of overlying layers (**Figure 3.2**; Cuffey and Paterson, 2010). As the glacier grows, gravity causes it to flow downslope. This movement occurs through two main processes: (1) plastic deformation of the ice, driven by shear stress that increases with ice thickness and slope, and (2) basal sliding, where the glacier slides over its bed, often facilitated by meltwater at the base (Hooke, 2019). At the glacier terminus, known as the *ablation zone*, ice is lost through melting, sublimation, or calving. The *Equilibrium Line Altitude (ELA)* marks the boundary between the accumulation zone, where ice accumulates and the ablation zone, where ice loss occurs (**Figure 3.5**).

Glacier mass balance refers to the net change in a glacier's mass over a specific period, typically measured annually. It is determined by the difference between net accumulation (gain of ice and snow) and ablation (loss of ice and snow). Accumulation primarily occurs through snowfall, while ablation encompasses processes such as melting, sublimation, and calving. A net positive

(+) mass balance indicates that a glacier is gaining mass, leading to potential advance, whereas a net negative (-) mass balance signifies mass loss and potential retreat. Monitoring mass balance is crucial for understanding a glacier's health and its response to climatic variations (Rounce et al., 2023). Factors influencing mass balance include mean annual and decadal temperatures, precipitation patterns, and surface albedo (Hooke, 2019; Steiner et al., 2015; Zemp et al., 2015). Increases in glacier mass loss contribute directly to global sea-level rise by adding freshwater to the oceans (Jacob et al., 2012; Zemp et al., 2019). Also, a net negative mass balance specifically in high mountain regions can reduce downstream water availability (i.e. lake ecosystems, agriculture, hydropower production, etc; Andres et al., 2018; Beniston and Stoffel, 2014; Narro Pérez et al., 2023; Viviroli and Weingartner, 2004)

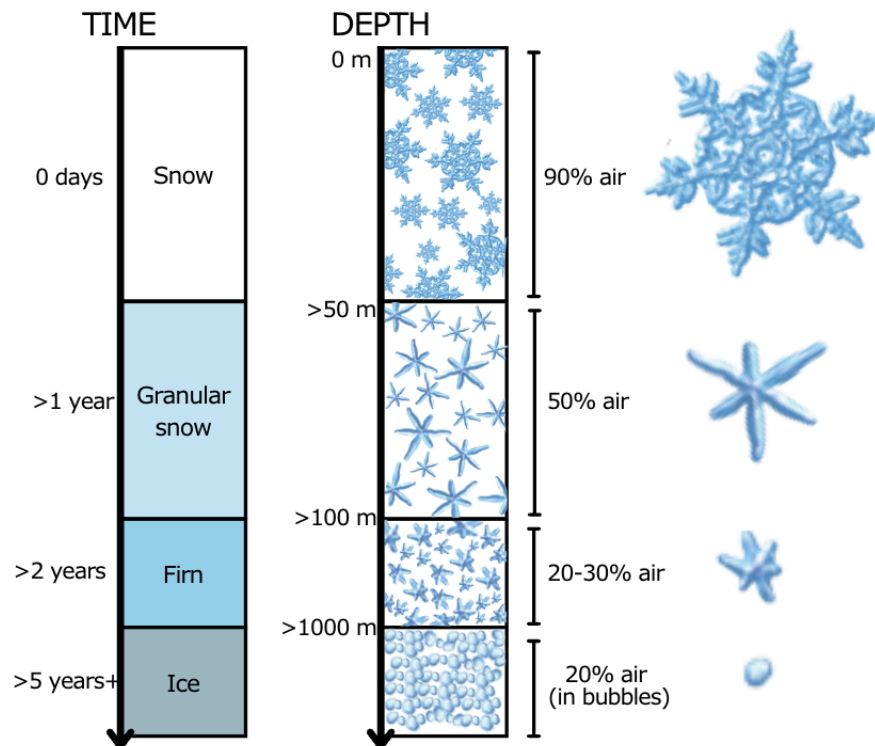


Figure 3.2. Schematic diagram of glacier ice formation. As time increases (depth not to scale), the compaction, mass, and pressure of water increases with less air bubbles. Redrawn by Chimira Andres and adapted from Mattavelli (2016).

As glaciers flow, they can erode the underlying rock through abrasion, scraping with embedded debris, and plucking, where meltwater loosens, lifts, and dislodges masses of bedrock. These processes shape landscapes with glacial landforms such as moraines, eskers, and kettle-kame terraces (more on Section 3.1.3; Benn and Evans, 2014; Eyles et al., 1985). As climate changes on Earth, glacial landsystems in the cryosphere become increasingly susceptible, making glaciers key indicators of environmental change (IPCC, 2019).

3.1.2.1 Thermal regimes of glaciers: *How does temperature affect glaciers?*

The thermal regime of a glacier is a function of ice temperature (function of climate, internal friction, and geothermal heat flow) and the pressure of the ice (ice thickness; Cuffey and Paterson, 2010). Glaciers are classified based on their thermal regimes (**Figure 3.3**), which influence their movement and erosional capabilities: (1) *Warm- (or wet) based glaciers* (also known as temperate glaciers) have ice at or near the pressure melting point, which is the temperature at which ice melts under a given pressure (i.e. decreases as pressure increases) throughout their mass (Paterson, 1994).

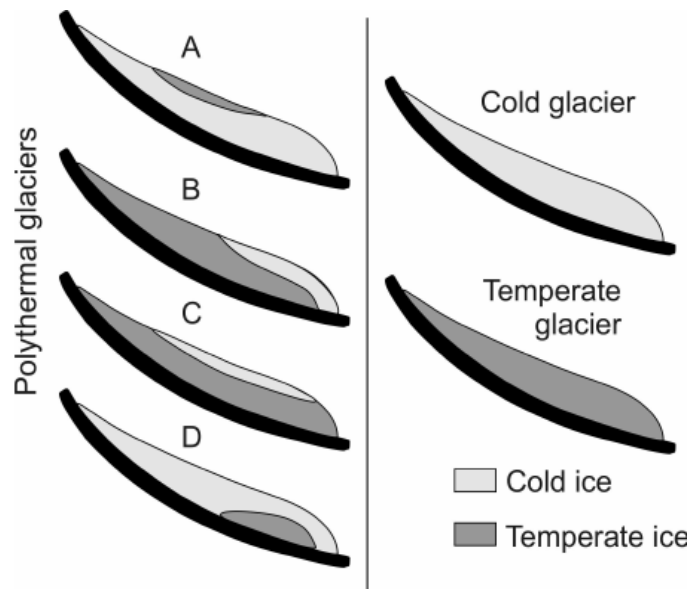


Figure 3.3. The three thermal structures of glaciers. The left panel shows variations of glaciers with polythermal regimes and the right panel distinctly categorizes wet-based and cold-based glaciers. Adapted from Dobiński (2012) and Blatter (1990).

The amount of meltwater at the base of a glacier controls glacial velocities, erosion, transfer, and deposition of debris, as well as being an important factor in controlling ice deformation (Benn and Evans, 2014; Evans and Gooster, 2014). Glacial sediments and landforms vary widely between different thermal regimes. The presence of liquid water at their base acts as a lubricant, facilitating basal sliding. This results in relatively rapid movement and erosional activity compared to cold-based glaciers, which lack basal meltwater and therefore move more slowly and erode minimally. Warm-based glaciers are typically found in temperate regions with high snowfall and moderate temperatures (Clarke et al., 1984). (2) *Cold- (or dry) based glaciers* have ice temperatures below the pressure melting point, causing them to be frozen to their beds. Consequently, they exhibit minimal basal sliding and erosion, with ice motion primarily occurring through internal (viscous) deformation of the ice. Viscous deformation refers to the slow, internal flow of ice under its own weight in response to stress. In glaciers, this happens when stress from the overlying ice causes the ice crystals and lattice to slowly change shape and move, even though the ice remains solid (Cuffey and Paterson, 2010). Unlike basal sliding,

which involves the glacier sliding over liquid water at its base, viscous deformation is entirely internal and does not require meltwater. Cold-based glaciers are common in polar regions on Earth and Mars, where persistently low temperatures keep the ice frozen to its bed. On Mars, the lack of depositional and erosional landforms typically associated with basal meltwater further supports this interpretation (more on Section 3.2; Garvin et al., 2006; Head and Marchant, 2003; Waller, 2001). (3) *Polythermal glaciers* exhibit characteristics of both warm-based and cold-based glaciers. Polythermal glaciers have zones of temperate ice, usually in thicker sections where insulation allows the ice to reach the pressure melting point, and cold ice, typically near the surface and margins. This results in complex flow dynamics and erosional patterns (Aschwanden and Blatter, 2009; Blatter and Hutter, 1991; Irvine-Fynn et al., 2011).

3.1.2.2 Glacial landforms and environments

Glaciers are powerful geomorphic agents that shape landscapes through erosion and deposition. Erosional processes primarily include *plucking* and *abrasion*. Plucking occurs when the glacier freezes onto bedrock and pulls away fragments as it moves, while abrasion results from debris embedded in the ice scouring and polishing the underlying surface. These processes carve classic glacial erosional landforms (Evans and Gooster, 2014).

U-shaped valleys are steep-sided, flat-bottomed troughs created as glaciers erode and widen pre-existing river valleys. Often, these valleys will also have *hanging valleys*, which are smaller tributary valleys that were eroded less deeply than the main U-shaped valley (Benn and Evans, 2014; Hirano and Aniya, 1988). They are left perched above it and commonly marked by waterfalls (on Earth) at their junction. *Cirques* are bowl-shaped depressions formed and shaped at the head of a glacier by subglacial erosion (Evans, 1977). *Arêtes* are sharp ridges that form between adjacent cirques or glacial valleys, sculpted by erosion from ice on either side (Barr and Spagnolo, 2015). *Overdeepened basins* are bedrock depressions carved by subglacial erosion, often forming beneath thick or fast-flowing ice (Evans and Gooster, 2014; Preusser et al., 2010). Overdeepened basins and hanging valleys will be two of the key landforms of interest in Chapter 4.

As glaciers retreat or stagnate, they also deposit unsorted sediments known as glacial till and glacial diamict. Glacial diamict or glacial till refers to a heterogeneous mixture of sediment grain sizes such as clay, silt, sand, gravel, and boulders deposited directly by ice without the influence of flowing water (Eyles et al., 1983). This results in a variety of depositional landforms (**Figure 3.4**).

Moraines are ridges of very unsorted till deposited along glacier margins and are among the most prominent glacial landforms. There are several moraine types and morphologies. For example, lateral moraines form along glacier sides, medial moraines form along the center of a glacier (can also form where lateral moraines from two merging glaciers meet), while terminal moraines mark the maximum extent of advance, and ground moraines blanket the terrain beneath retreating ice. Their morphology and spatial distribution provide valuable records of past glacier positions and dynamics. *Drumlins* are streamlined, teardrop-shaped hills formed beneath glaciers by the deformation of subglacial sediments or sculpting of bedrock, aligned parallel to ice flow. Their elongated shape indicates the direction of past glacier movement, and fields of drumlins often occur in formerly glaciated lowlands. *Eskers* are sinuous ridges of sand and gravel ridges deposited by subglacial and/or englacial meltwater streams (Boulton et al., 2009; Shreve, 1985). They trace the paths of drainage systems and can extend for tens of kilometers, preserving evidence of hydrologic processes beneath/within the ice.

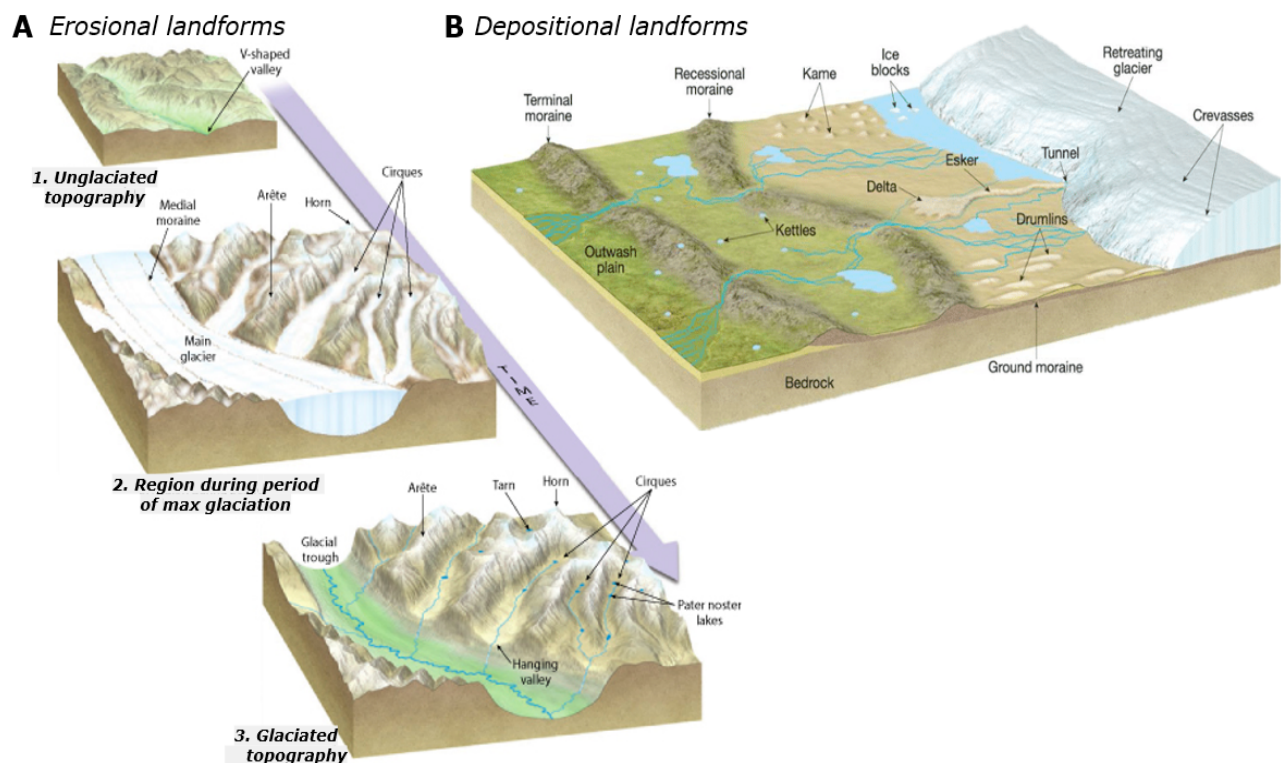


Figure 3.4. Principal (A) erosional and (B) depositional landforms as a result of past deglaciation. Adapted from (Benn and Evans, 2014).

Beyond these landforms, glaciers also operate within distinct zones and environments where sediment transport and deposition occur (**Figure 3.5**). The diverse glacial geomorphology and

formation of these landforms occur in environments that are broadly categorized into within four (4) main environments: subglacial, englacial, supraglacial, and paraglacial:

- *Subglacial environments*: Found beneath active ice, these are dominated by till generated through lodgement, melt-out, or deformation (Boulton et al., 2009; Clarke et al., 1984). Subglacial deformation is now considered one of the most important processes in till formation, producing massive, structureless diamict that cover vast areas, such as those seen across Southern Ontario, Canada (Eyles et al., 1985; Mahaney et al., 2014).
- *Englacial environments*: Are within the glacier body itself. Sediment is transported inside the ice via incorporation during basal freezing or from surface debris that becomes buried. When the ice melts, this debris is released and contributes to depositional features such as medial moraines (Gulley et al., 2009).
- *Supraglacial environments*: Located on the surface of the glacier including the front of the ice margin, these areas accumulate debris from rockfall, landslides or englacial transport (Schomacker and Benediktsson, 2018). As the glacier melts, sediment is released and deposited directly onto the landscape in chaotic, unsorted masses often as moraines. Notably, this environment is the most relevant to debris-covered glaciers on Mars discussed in this thesis in Chapters 4 and 5.
- *Paraglacial environments*: These develop after glacier retreat, when formerly glaciated landscapes adjust to the removal of ice. The destabilization of sediment and slopes leads to intensified mass wasting, fluvial reworking, and slope failure. Paraglacial environments are characterized by features such as alluvial fans, reworked moraines, and debris flows, and they play a significant role in the landscape's transition from glacial to post-glacial conditions (Ballantyne, 2002).

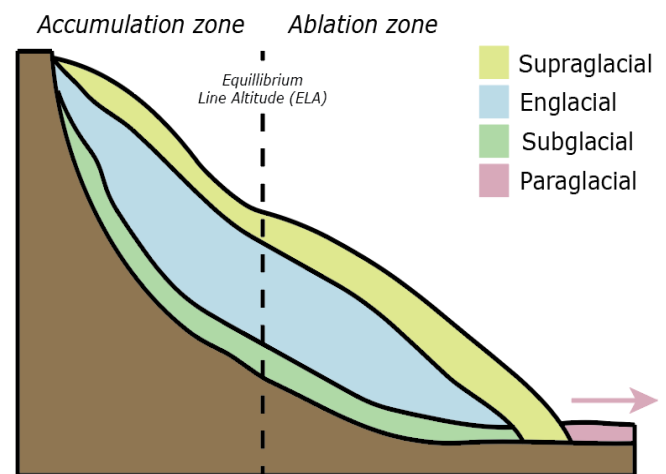


Figure 3.5. Anatomy of a glacier showing simplified glacier zones and environments.
Drawn by Chimira Andres.

In some cold climate settings, periglacial environments (detailed in Section 3.1.3) also contribute to the sedimentary record. These areas, which may lie adjacent to glaciated regions, experience freeze-thaw cycles, frost heaving, and permafrost-related processes. Although not directly

glacial, periglacial deposits such as ice-wedge casts and patterned ground, preserve important evidence of paleoclimatic conditions (Ballantyne, 2018; French, 2017).

3.1.2.3 Debris-covered glaciers

Debris-covered glaciers are characterized by a surface mantle/layer of rock and sediment, which can substantially alter melt rates and glacier behavior. The thermal impact of this debris is dependent on its thickness. Thin debris layers can enhance ablation by lowering surface albedo, increasing the absorption of solar radiation and accelerating melting, while thick debris layers act as insulating blankets that reduce heat flux and protect the underlying ice from melting (Evatt et al., 2015; Nicholson and Benn, 2013, 2013; Østrem and Brugman, 1966). The distribution and origin of surface debris are influenced by surrounding topography, rockfall activity, and internal glacier dynamics (Kirkbride and Deline, 2013). Material may be delivered supraglacially from valley walls via rockfall, avalanching, or debris sloughing, which deposits sediment directly onto the glacier surface. It may also be transported englacially, where debris is buried within the ice due to deformation, basal freeze-on processes, or incorporation at the glacier bed, and later re-emerges as the glacier flows and melts (Bigelow et al., 2020; Gulley et al., 2009).

Debris-covered glaciers often develop in regions of high relief where sediment availability is high and climatic conditions permit prolonged ice retention under a protective debris layer (**Figure 3.6**; Scherler et al., 2018). Their subdued surface morphology and insulating properties complicate mass balance measurements and make them more resilient to short-term climate warming compared to “clean-ice” glaciers. However, they may also contain large volumes of stored meltwater and can contribute to hazards such as glacial lake outburst floods (GLOFs) if moraine-dammed lakes with large volumes of water form at their margins (Richardson and Reynolds, 2000). Understanding debris-covered glaciers is vital because their insulating debris layers can preserve ice long after climatic conditions become unfavorable for clean-ice glaciers, providing evidence of former glacier extent and dynamics.

This glacier type is the current hypothesis and direct Earth analogue for glaciers on Mars, which are the focus of Chapters 4 and 5. Additionally, changes in debris thickness and distribution influence melt (or sublimation) patterns, making these glaciers key indicators of how ice, with high debris input, responds to shifting climate and sediment supply. Beyond the limits of glacial ice, adjacent cold-climate regions are shaped not by flowing ice but by intense freeze-thaw

processes and the presence of permafrost. These periglacial domains play a critical role in cold-region geomorphology and often act as transitional zones between glaciated terrains and more temperate landscapes (French, 2017).

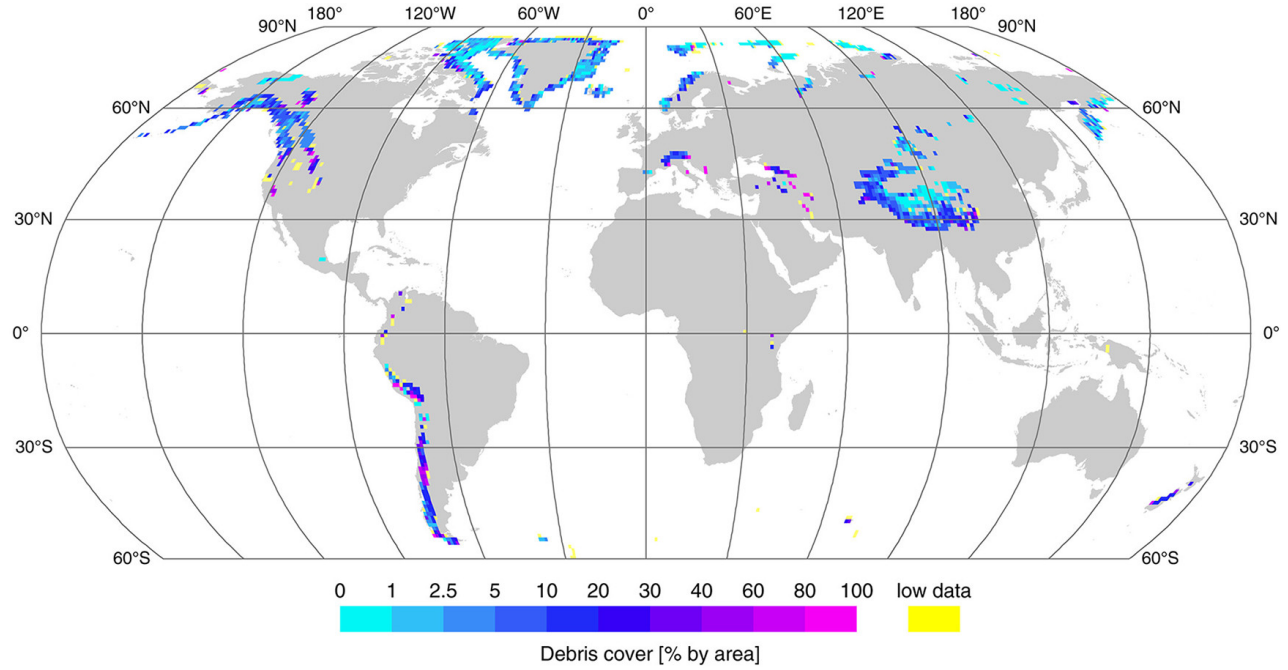


Figure 3.6. Debris coverage on glaciers around the world given as % by area and estimated from Landsat-8 satellite imagery between 2013 and 2015 with data binned by area with a tile size of $1^\circ \times 1^\circ$. From Scherler et al., (2018).

3.1.3 Periglacial domain: *How does subsurface ice form?*

The periglacial domain refers to cold-climate regions adjacent to glaciers and ice sheets where permafrost and freeze-thaw processes dominate the landscape. These environments are not glaciated but experience temperatures low enough to support long-term ground ice. Repeated freezing and thawing cycles lead to the formation of distinctive features such as patterned ground (detailed in Section 3.1.3.2.2; Ballantyne, 2018; Mackay, 1999, 1995). Another hallmark of periglacial environments is thermokarst terrain, characterized by irregular subsidence caused by the melting of excess ground ice and drainage (detailed in Section 3.1.3.2.1; Bouchard et al., 2020; Ulrich et al., 2010).

A ubiquitous landform to all periglacial landscapes in the cryosphere is the presence of *permafrost*, permanently frozen ground that holds various forms of subsurface ice and controls the formation and stability of surface features. Permafrost is present both on Earth and on Mars

and the following section will provide key details on the different types of ground ice and briefly describe mechanisms of formation of relevant permafrost landforms.

3.1.3.1 Permafrost

Permafrost is defined as ground (i.e. soil or rock, including ice and organic material) that remains at or below 0°C for at least two consecutive years (Péwé, 1969). It is a defining feature of the periglacial environment and underlies vast regions of the Arctic, subarctic, and high mountain areas (O'Neill et al., 2019; **Figure 3.7**). The presence and stability of permafrost are influenced by factors such as mean annual air temperature, snow cover, vegetation, soil composition, and hydrology (Smith, 1990; S. L. Smith et al., 2022). Permafrost is typically overlain by an active layer that thaws and refreezes seasonally. Changes in the thickness of this active layer serve as indicators of climate variability and permafrost degradation (Romanovsky et al., 2010). As climate warming accelerates, widespread thaw of permafrost is occurring, leading to surface subsidence (thermokarst), increased greenhouse gas emissions, and damage to infrastructure (Smith, 1990; Smith and Riseborough, 2002).

Permafrost contains various forms of ground ice, which differ in origin and impact on landscape stability. *Pore ice* forms within the voids between soil grains and is the most common, though generally low in volume. Pore ice typically forms

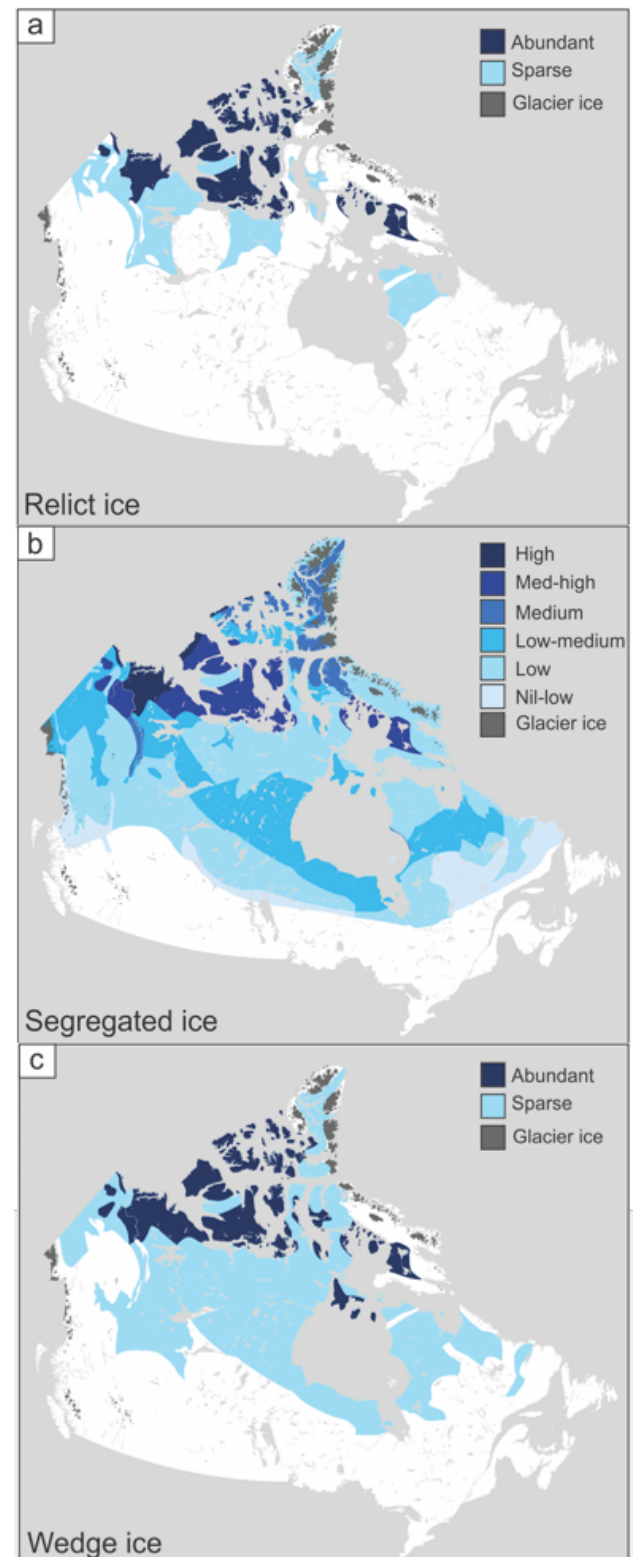


Figure 3.7. The Permafrost map of Canada. (a) relict (“massive”) ice, (b) segregated ice, and (c) wedge ice (as modelled and mapped by O'Neill et al., 2019). These maps were also originally modified from Heginbottom et al., 1995.

when liquid water (i.e. either present in the soil, drawn upward by capillary action, or infiltrated from surface melt) freezes within the pore spaces of moist, fine-grained sediments during subzero conditions (French and Shur, 2010). However due to the absence of liquid water on present-day Mars, pore ice is hypothesized to form primarily through vapour diffusion, where atmospheric water vapour migrates into the regolith and freezes under cold surface conditions (Clifford, 1991; Hudson et al., 2007; Mellon and Jakosky, 1993). When the volume of ice exceeds the sediment's pore space, it is referred to as *Excess ice*, which can also include segregated ice, ice wedges, or massive ice bodies (French and Shur, 2010; French, 2017). *Segregated ice* forms when water migrates to freezing fronts and accumulates, often creating discrete *ice lenses* parallel to the ground surface in fine-grained, frost-susceptible soils (**Figure 3.7**; Murton and French, 1994). *Ice wedges* develop when thermal contraction cracks are filled by meltwater that freezes and expands over successive winters, forming large, wedge-shaped bodies (Mackay, 1995, 1990; Mackay and Burn, 2002). *Massive relict ice*, such as buried glacier ice, represents remnants of past glacial periods and can persist beneath sediment cover for thousands of years (Murton, 2005; Murton et al., 2005).

These ice types affect ground stability by controlling the mechanical strength and volume of permafrost soils. When ground ice thaws (especially excess ice) it leads to subsidence and surface collapse, which reshape the landscape and disrupt drainage patterns. Remote sensing, ground-based geophysical surveys, and borehole monitoring are key tools for assessing permafrost extent, thermal state, and ice content in both Earth and planetary analogue environments (Carr and Schaber, 1977; Dobinski, 2011; Kreslavsky et al., 2008).

3.1.3.2 Periglacial landforms

Periglacial environments host a diverse range of landforms shaped by freeze-thaw processes, ground ice dynamics, and permafrost degradation (Ballantyne, 2018; Harris and Murton, 2005; Washburn, 1980). Landforms outlined in this section, particularly, *thaw slumps* and *polygonal terrain* are especially relevant for Chapter 7 as they are the two target landforms for geophysical fieldwork in the Canadian Arctic (**Figure 3.8**). *Rock glaciers* are highlighted in Section 3.1.3.2.3 as a landform of interest as they are also an alternate analogue for Viscous Flow Features (VFFs) on Mars. Understanding their formation and distribution is essential for interpreting open questions in current and past climate, terrain stability, and landscape evolution not only on Earth, but also on Mars.

3.1.3.2.1 Thermokarst and Retrogressive Thaw Slumps (RTS)

Thermokarst refers to land surface dominated by the thawing of ice-rich permafrost producing irregular marshy hummocks (French, 2017; **Figure 3.8A**). When excess ground ice melts, the resulting volume loss leads to subsidence, forming irregular terrain such as depressions, sinkholes, collapsed scarps, and thaw lakes (Grosse et al., 2013; Kokelj and Jorgenson, 2013). Thermokarst features can develop rapidly and vary widely in scale, depending on ice content, sediment type, and thermal regime (Grosse et al., 2013). Their formation alters surface hydrology and can disrupt vegetation, infrastructure, and carbon storage, and can also be a key indicator of permafrost degradation and thermal instability (Burn et al., 2015; Schuur and Mack, 2018). Retrogressive thaw slumps are caused by thawing of ice-rich permafrost and subsequent melting of ground ice on slopes (Nesterova et al., 2024).

A specific form of thermokarst is the Retrogressive Thaw Slump (RTS; **Figure 3.8B**). These features form when ice-rich headwalls become exposed, typically by erosion or vegetation loss, triggering melt and slope failure (Lewkowicz and Way, 2019). As thawing continues, the slump expands upslope (retrogresses), creating a steep headwall, a muddy or hummocky floor, and lobes of displaced debris at the base. The evolution of RTS is controlled by factors such as solar radiation, precipitation, slope angle, and sediment composition (Li et al., 2024; Nesterova et al., 2024). Both thermokarst and RTS are critical to monitor due to their rapid expansion and impact on terrain stability, hydrological systems, and feedbacks to global carbon cycles. Their morphology and development can be effectively tracked using time-series satellite imagery, digital elevation models, and ground-based observations (Kokelj and Jorgenson, 2013; Nesterova et al., 2024; Van Huissteden et al., 2021). Thaw slumps will be one of the main landforms of interest in Chapter 6.

3.1.3.2.2 Polygonal ground

Polygonal ground is a common and diagnostic periglacial landform formed by thermal contraction cracking in ice-rich soils (Washburn, 1956; **Figure 3.8C**). These cracks can fill with ice or sediment over time, resulting in visible polygon-shaped surface patterns. There are two main types: ice-wedge polygons, which form when water enters cracks and freezes, creating large vertical wedges of ice; and sand-wedge polygons, which form in drier conditions with less ice (French, 2017). Polygon development is sensitive to surface temperature, moisture, and

permafrost stability, making them useful indicators of recent and historical freeze-thaw activity (Mackay, 1980). Polygons typically range from ~5-30 m in diameter and exhibit a variety of configurations, including high-center, low-center, and flat-center morphologies, often arranged in honeycomb-like networks (i.e. rectangular, pentagonal or octagonal shapes). Their geometry and degradation state are often used to infer climatic history and ground ice content in both Arctic and Martian settings (Ballantyne, 2018; Burn, 1998; French, 2017; Levy et al., 2009; Séjourné et al., 2011). Polygonal patterned ground will be one of the main landforms of interest in Chapter 6 of this thesis.

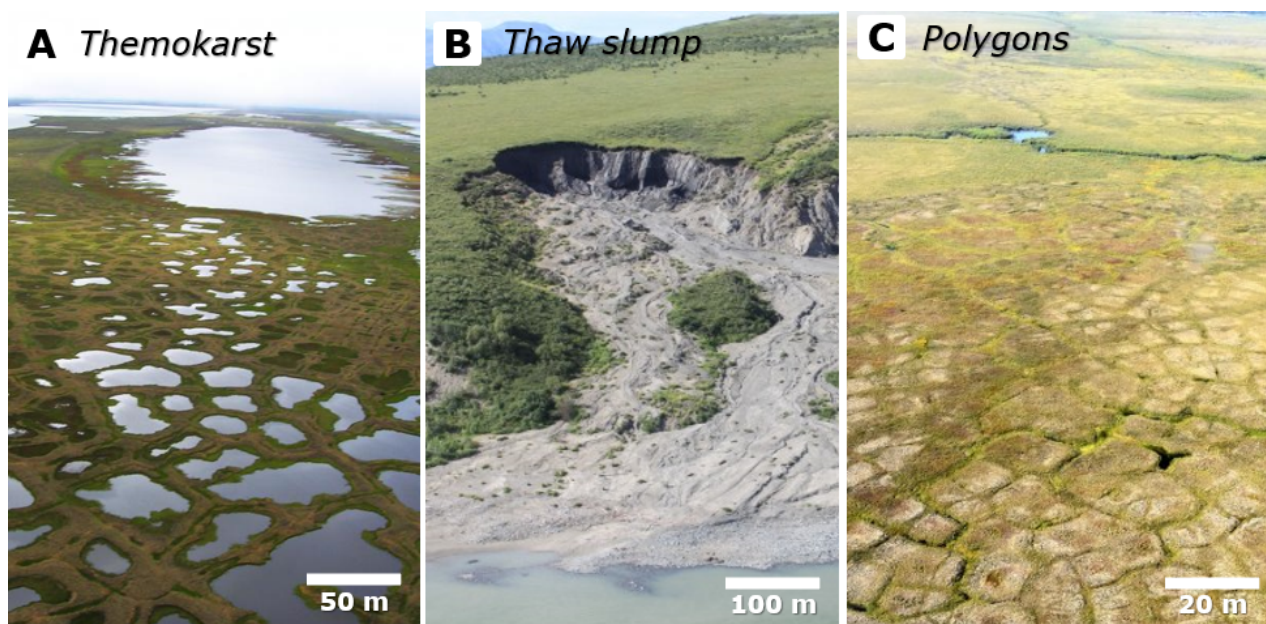


Figure 3.8. (A) An aerial view of arctic thermokarst lakes. Photo by University of Alaska Fairbank (UAF) (B) Retrogressive thaw slump on the Noatak river; the top of the slump is ~270 m above river water level. Photo by Rory Nicols (C) Ponds form over polygonal terrain, a distinctive permafrost landform. Photo by the Government of the Northwest Territories, NTGS/NWT Centre for Geomatics

3.1.3.2.3 Rock glaciers

Rock glaciers are lobate or tongue-shaped landforms composed of a mixture of angular rock debris and interstitial ice, typically forming in high-relief, periglacial environments (Giardino et al., 1987; Humlum, 1998; Jones et al., 2019). These features deform and flow downslope slowly, typically centimeters to meters per year, through a combination of ice creep, shear deformation, and frost-related processes such as frost heave (Haeberli et al., 2006). Their internal structure often includes interstitial ice, segregated ice lenses, or massive relict glacial ice, all protected beneath a thick, insulating debris that inhibits surface melt (**Figure 3.9; Figure 3.10;** (Berthling, 2011).

There are several formation hypotheses for rock glaciers: (*Periglacial*) *rock glaciers* form through the creep of ice-rich permafrost in talus or colluvial slopes, where moisture migrates toward freezing fronts, forming segregated ice lenses (Giardino and Vitek, 1988). *Glacial-derived rock glaciers* evolve from debris-covered glaciers as the surface ice melts and degrades, leaving behind buried glacier ice encapsulated by thick debris (Anderson et al., 2018). *Composite or hybrid rock glaciers* result from interactions between both processes, often involving the relict cores of former glaciers overlain by periglacially derived debris and ice (Haeberli et al., 2006; Meng et al., 2023; Berthling, 2011)

These landforms occupy an intermediate space between active glaciers and periglacial slopes, governed by the balance of ice activity and debris input (**Figure 3.9A**). As debris input increases and glacial activity decreases, landforms evolve from debris-covered glacier snouts to ice-cored moraines, glaciogenic lobes, and eventually rock glaciers. Their surfaces often display transverse ridges, furrows, and other indicators of long-term, plastic deformation (Haeberli et al., 2006; Hamilton and Whalley, 1995; **Figure 3.9B**).

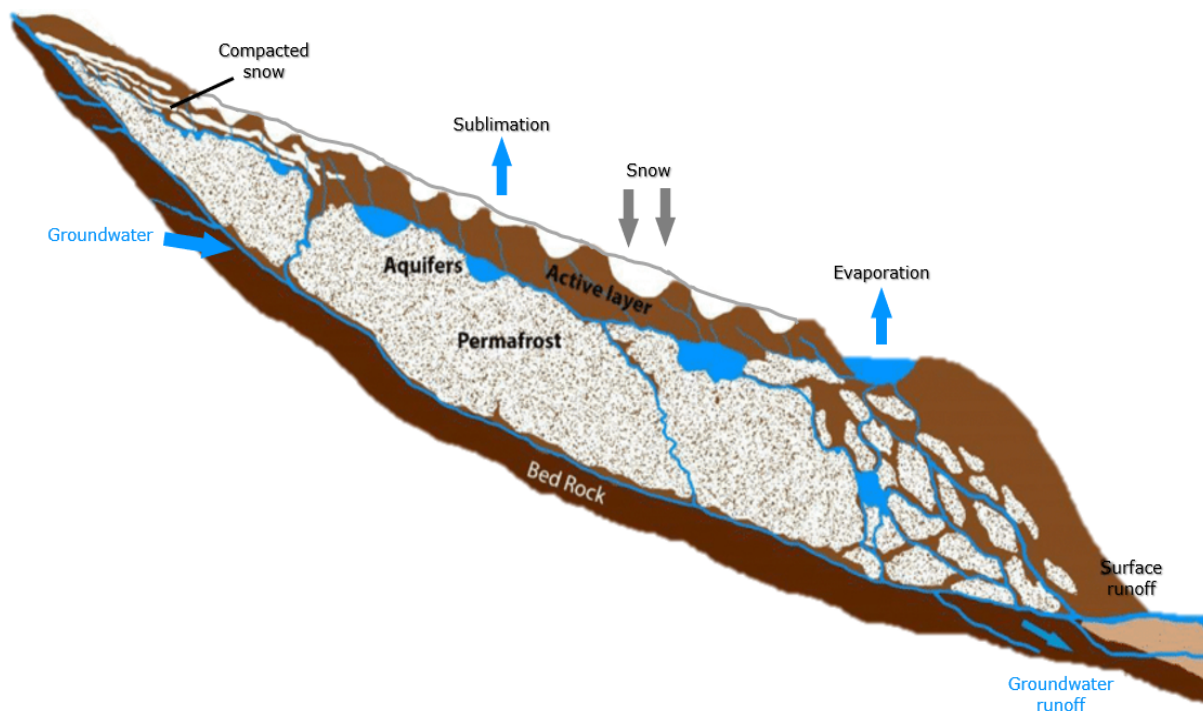


Figure 3.9. Conceptual cross-section of a rock glacier showing key components including the active layer, permafrost core, and internal aquifers. Snowfall and groundwater recharge feed the system, while outputs include sublimation, evaporation, surface runoff, and groundwater discharge. This highlights the interplay between permafrost, seasonal thaw, and hydrological flow in cold mountain environments. Modified from Burger et al. (1999) and Giardino et al. (1992).

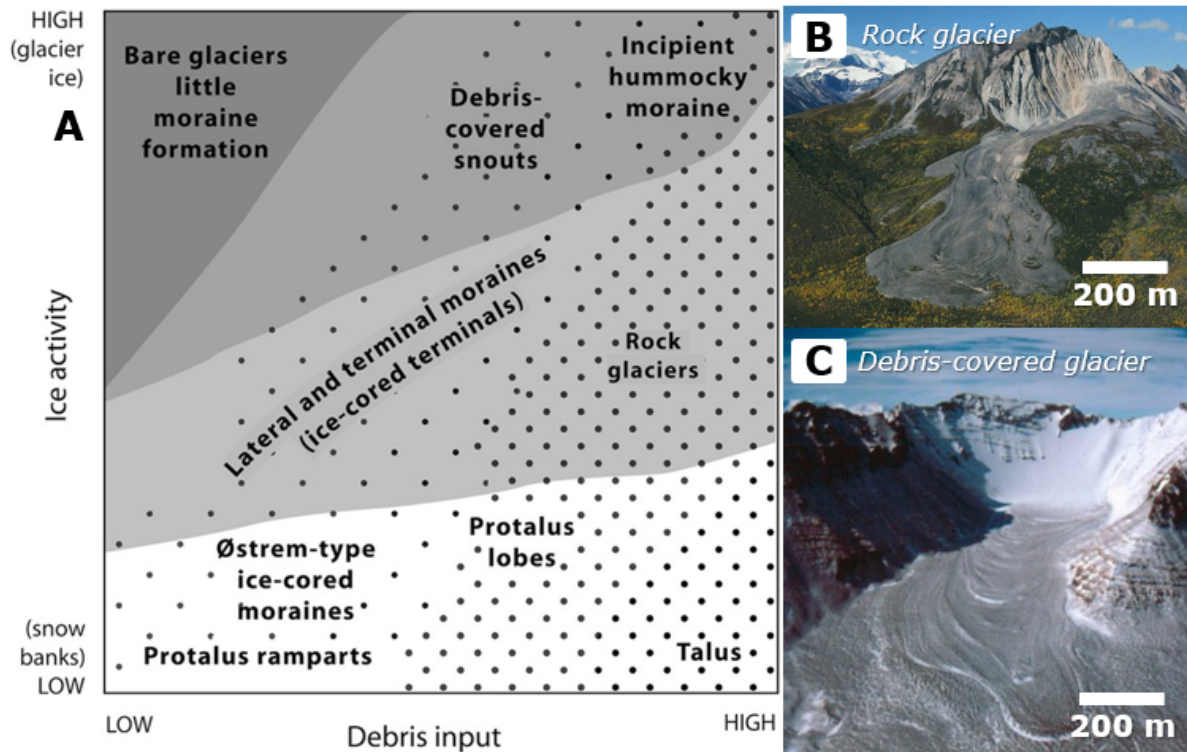


Figure 3.10. Periglacial and glacial landforms can be evaluated as a function of ice activity and debris input. Dots indicate relative debris content and shading broadly groups glacier types. Ice activity and debris input do not have to be constant in time, and these types are not distinct (but can be distinguished in many regions). Considering any given deposit as a function of debris and ice sources can help to identify formation processes as a function of climate controls. From Whalley (2009). **(B)** An aerial view of Sourdough Rock Glacier in Alaska (Source: NPS Natural Resources). **(C)** An aerial view of Mullins Glacier in Antarctica (Photo by Dr. Joseph Levy).

While *debris-covered glaciers* and *rock glaciers* can appear morphologically similar, both featuring lobate forms, surface ridges, and insulating debris covers, their internal structure and deformation mechanisms differ (**Table 3.1**). Debris-covered glaciers typically consist of continuous glacier ice overlain by a few centimeters (>2 cm) to meters (~1-3 m) of surface debris layer (thickest at the terminus), and their motion is primarily driven by basal sliding and internal ice deformation (Benn et al., 2012; Quincey et al., 2009), with typical flow rates of ~30-100+ m per year (Heimsath and McGlynn, 2008; Sam et al., 2016; **Figure 3.9C**). In fact, some debris-covered glaciers can transition into *surge-type* glaciers, undergoing short-lived episodes of rapid flow (hundreds of meters to several kilometers per year), often due to changes in subglacial hydrology or mass redistribution (Benn and Evans, 2014; Clarke et al., 1984; Fowler et al., 2001).

On the other hand, rock glaciers are mostly composed of a coarse debris matrix with a heterogeneous mix of interstitial, segregated, and/or relict glacial ice within the sediment grains. They deform more slowly, through a combination of ice creep within a debris-ice matrix and

freeze-thaw processes, similar to permafrost (**Figure 3.9B**) with rates of 0.1-1 m per year, although inactive or relict rock glaciers move slower or not at all (Berthling, 2011; Haeberli et al., 2006; Hu et al., 2023; Kääb et al., 2003).

Table 3.1 Comparison of debris-covered and rock glacier properties.

Properties	Debris-Covered Glacier	Rock Glacier
Approximate Ice Content (%)	80-95% ice 5-20% debris	40-60% ice 40-60% debris
Debris Thickness	Few cm to >1 m	The debris is the matrix, not just surface
Internal Structure	Continuous glacier ice under debris	Ice-debris matrix with interstitial, segregated, or buried glacier ice
Flow Mechanism	Internal ice deformation and basal sliding	Creep of ice-debris mixture and freeze-thaw processes
Flow Rate	~10-300 m/year (can exceed during surges)	~0.01-1 m/year
Surface Features	Lobate forms, supraglacial ponds, ice cliffs, kame-kettle topography	Lobate forms, surface ridges, furrows
Formation Environment	<i>Glacial</i> (alpine glaciers with debris input)	<i>Periglacial</i> or <i>paraglacial</i> (deglaciated valleys, cold, high-relief environments)

However, the distinction between debris-covered glaciers and rock glaciers is not always clear. Transitional landforms exist, where glacier-derived ice is buried beneath thick debris and begins to deform under permafrost conditions, resembling the internal structure and kinematics of rock glaciers (Berthling, 2011; Monnier and Kinnard, 2017). In these hybrid systems, the boundary between glacial and periglacial processes blurs, especially in high-relief regions where glacier ice can become relict and permafrost conditions stabilize the landform. Previous literature have proposed the term "ice-cored rock glacier" or "glaciogenic rock glacier" to describe such features that inherit ice from glaciers but evolve under permafrost conditions (Johnson, 1980; Martin and Whalley, 1987; Potter Jr, 1972; White, 1976).

Understanding this glacial–periglacial continuum is critical for planetary analogue studies, particularly on Mars, where VFFs may span both glacial-periglacial domains. Radar and morphometric data often do not resolve internal structure, making it difficult to distinguish between debris-covered glacier and rock glacier analogues from surface features alone. This ambiguity reinforces the importance of Earth analogues and the need for combined structural, thermal, and compositional datasets.

3.2 Ice on Mars

This section provides the current the state of knowledge of the cryosphere on Mars including early investigations of ice on Mars and current hypothesis of ice transport from the poles to the midlatitude regions. This is then followed by the descriptions of specific landforms associated with buried ice. Emphasis is placed on viscous flow features (VFFs), interpreted as Martian analogues of debris-covered glaciers, and the latitude dependent mantle (LDM) due to their significance for understanding climate-driven ice accumulation, landscape evolution, and potential resources for in-situ resource utilization (ISRU). Section 3.2 concludes by discussing insights from orbital observations, theoretical models (i.e. Subsurface Water-ice Mapping team's work), and analogue comparisons to cold regions on Earth discussed in the previous Section 3.1 *Ice on Earth*.

Water-ice is a major component of the Martian near-surface environment, especially within the poles to the mid-latitudes (30° – 60° in both north and south hemispheres). It is thought to occur in various forms such as pore ice (Aharonson and Schorghofer, 2006; Mellon et al., 2008; Mellon and Jakosky, 1993), interstitial ice (Mellon et al., 1997), and excess ice, which includes ice lenses or massive buried layers that exceed the pore volume of the regolith (Bramson et al., 2015; Séjourné et al., 2012; Soare et al., 2015). These ice types are typically preserved beneath a thin insulating layer of dust or debris (<15 m), which protects them from sublimation under Mars' cold, low-pressure atmospheric conditions (Baker and Carter, 2019; Conway and Balme, 2014). This insulating debris has been key to preserving buried ice for millions of years in a sublimation-prone environment (detailed in Section 3.2.8). Theoretical models and remote sensing observations have shown that this excess ice can persist for millions of years if shielded by even a few decimeters of surface material (Dundas et al., 2023, 2010; Fastook et al., 2011). In addition to the north and south polar caps, Mars also hosts seasonal CO_2 and H_2O frost deposits that form and disappear seasonally in response to temperature changes, and perennially exposed surface ice at both polar caps, where layered deposits of water and carbon dioxide ice have accumulated over millions of years (Jakosky, 1985; Kieffer and Titus, 2001; Piqueux et al., 2015). Understanding the origin, distribution, and stability of these ice forms is essential not only for reconstructing recent Martian climate cycles, but also for evaluating future in situ resource use for human exploration.

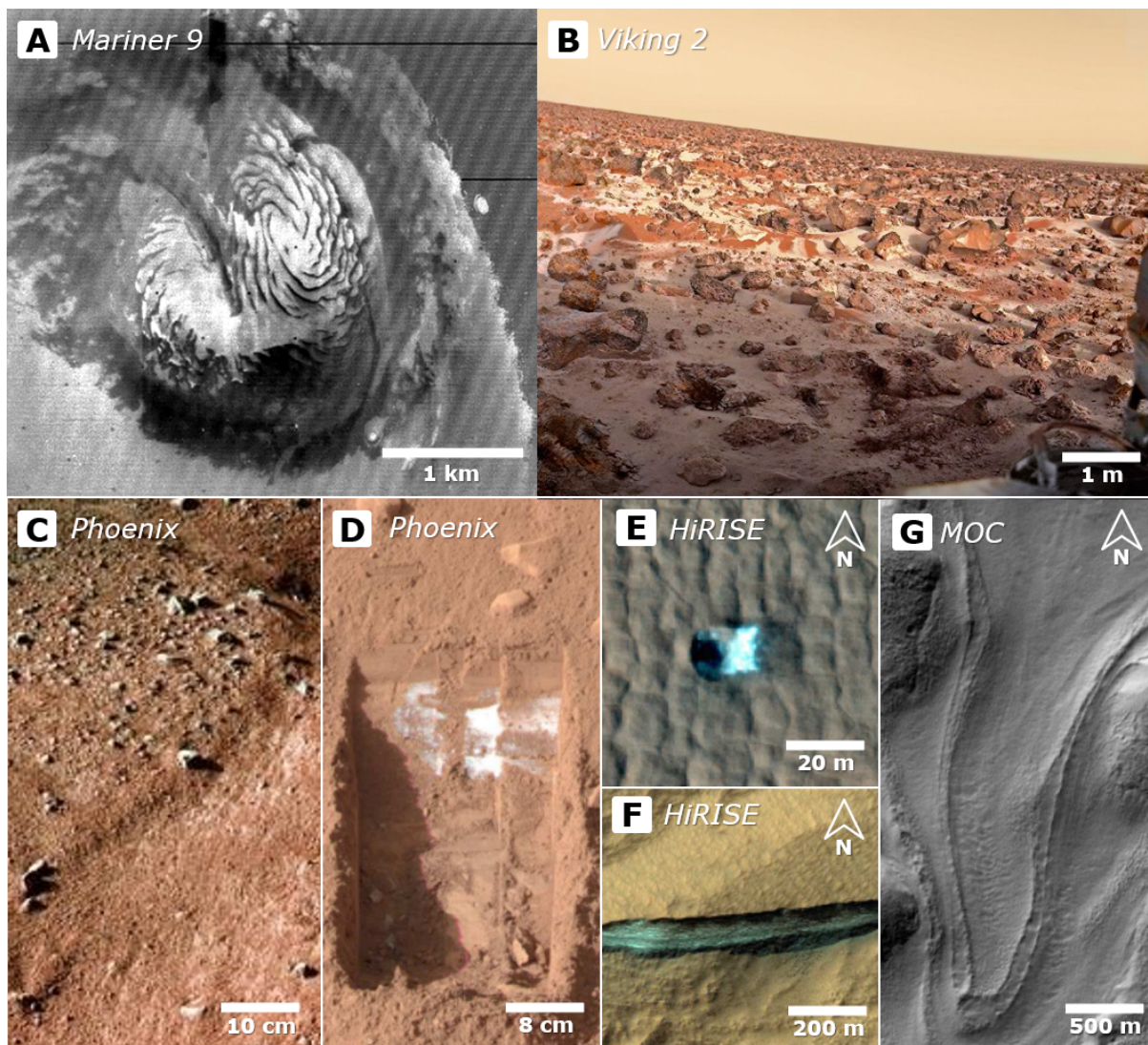


Figure 3.11. Key spacecraft observations of Martian ice and cryospheric features. (A) Mariner 9 image of the north polar cap, revealing spiral troughs in polar layered deposits. (B) Viking 2 lander view showing seasonal ground frost in the northern mid-latitudes. (C) and (D) Phoenix lander images of surface (C) shows polygonal troughs near the landing site and (D) shows a trench exposing subsurface water-ice in the northern plains, which is the first in-situ measurement of subsurface ice. (E) and (F) HiRISE images of recent impact craters and mid-latitude icy scarp revealing exposed buried ice (HiRISE Image IDs: ESP_011494_2265 and ESP_040772_1215). (G) Mars Orbiter Camera (MOC) image (Image ID: M1800897) of a lobate flow interpreted as a glacial landform. These observations demonstrate the widespread and diverse expressions of water-ice across Mars. *Image credits: Modified from Butcher, 2022. (A)-(F) NASA/JPL/University of Arizona/Texas A&M University/Malin Space Science Systems (G) Malin et al., 2001 Mars Orbiter Camera Image Gallery (http://www.msss.com/moc_gallery/).*

3.2.1 Early Investigations of Ice on Mars

The first evidence of ice on Mars was first inferred from the Mariner (1971) and Viking (1975) missions revealing spiral troughs in north polar layered deposits (NPLD; **Figure 3.11A**). The spiral troughs were initially interpreted as aeolian and features signifying ice layering (Howard, 2000, 1980, 1978). The presence of water-ice on Mars was further hypothesized and confirmed during the Viking 2 lander (1976) as evidenced from the seasonal ground frost on the landing site in

Utopia Planitia (47.6°N, 134.3°E) further suggesting shallow water-ice at mid to high latitudes (**Figure 3.11B**).

Initial spectral observations from orbit, including those by Hubble (late 1990s) and OMEGA (2004) on Mars Express, revealed widespread H₂O and CO₂ ice, as well as hydrated minerals that implied historical water-rock interaction (Appéré et al., 2011). Morphological evidence of past glaciation (i.e. such as VFFs) was also identified in Viking 2 imagery (Squyres, 1978; Lucchitta, 1984) and later confirmed with higher resolution data from Mars Global Surveyor (MGS) and MRO (**Figure 3.11G**). These features hinted at ice-flow processes but lacked direct compositional confirmation.

Table 3.2. List of space missions that contributed to the detection of ice on Mars.

Mission	Year Launched	Instrument(s)	Ice Discovery Highlights
Mariner 9	1971	Imaging System, Infrared Radiometer	First spacecraft to orbit Mars; provided detailed images of both the north and south polar caps.
Viking 2	1975	Imaging system	Observed potential ground ice features, ground frost forming seasonally in the northern mid-latitudes.
Hubble Space Telescope	1990	Wide Field and Planetary Camera	Observed seasonal changes in polar caps, supporting presence of CO ₂ and water-ice
Mars Global Surveyor (MGS)	1996	MOLA	Mapped the North Polar Layered Deposits (NPLD) and inferred their icy composition, first observations of terrain smoothing (ice-rich mantle)
Mars Odyssey	2001	Gamma Ray Spectrometer (GRS), THEMIS	Detected large subsurface hydrogen (interpreted as water-ice) at high latitudes, also revealed thermal behavior and properties of icy regions
Mars Express	2003	MARSIS Radar, OMEGA	Identified polar layered deposits and possible subsurface liquid water under south pole, also mapped water and CO ₂ ices and hydrated minerals
Mars Reconnaissance Orbiter (MRO)	2005	SHARAD, HiRISE, CTX, CRISM	Detected buried glaciers, ice-rich mantling, and stratified polar ice layers; spectral mapping of ice-related minerals
Phoenix Mars Lander	2007	Robotic Arm, TEGA	Directly observed and confirmed water-ice just below surface in northern plains
Trace Gas Orbiter (TGO)	2016	FREND (neutron spectrometer), NOMAD, CaSSIS	FREND mapped subsurface hydrogen; potential for ice mapping at finer scales

It was not until the early 2000s that more robust evidence emerged of a quantitative detection of subsurface water-ice advanced significantly with data from the Mars Odyssey Gamma Ray Spectrometer (GRS), which identified high hydrogen content poleward of ~45° latitude, interpreted as buried ice (Bandfield, 2007; Boynton et al., 2002). Thermal inertia data from THEMIS on the same mission further suggested icy mantling in dusty, high-latitude terrains. Direct confirmation came with the Phoenix Lander (2008), which excavated bright ice just centimeters below polygonal terrain at 68°N (Mellon et al., 2009; **Figure 3.11C** and **3.11D**). The excavated ice sublimated over

several days, confirming its water-ice composition. These on-ground observations were also supported by spectral data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) investigating a seasonal ice cycle specifically at the Phoenix landing site (68.2°N, 234.3°E; Cull et al., 2010b, 2010a) .

Subsequent observations from SHARAD (late 2000s - present) revealed strong radar reflections consistent with decameter-thick subsurface ice in multiple regions of Mars, such as in Deuteronilus Mensae (Petersen et al., 2018a; Plaut et al., 2009), Utopia Planitia (Stuurman et al., 2016), Arcadia Planitia (Bramson et al., 2015), and in the southern mid-latitudes (Holt et al., 2008). These detections were supported by numerical models of obliquity-driven climate change that predicted volatile migration and mid-latitude ice stability (Laskar et al., 2004; Madeleine et al., 2009).

More recently, HiRISE has imaged fresh impact craters and icy scarps exposing clean, bright subsurface (and layered) ice at mid-latitudes, often within 1 m of the surface (Dundas et al., 2023, 2018; Harish et al., 2020; Williams et al., 2022; **Figure 3.11E and 3.11F**). These findings confirmed that non-polar ice is not only present but also surprisingly shallow (Bandfield, 2007; Dundas et al., 2018; Mellon et al., 2004; Piqueux et al., 2019). In tandem with MRO, the Trace Gas Orbiter (TGO; 2016), has also advanced our ongoing understanding of Mars' cryosphere often planning coordinated mission campaigns for scientific targets. The FREND neutron spectrometer has presented a global map of Water Equivalent Hydrogen (WEH) in the upper meter of the Martian subsurface at an unprecedented resolution (Malakhov et al., 2022, 2020), while CaSSIS, the onboard high-resolution stereo imager, has provided spectral evidence for analyzing water-ice signatures (Rangarajan et al., 2024, 2022).

Together, these investigations demonstrate that ground ice on Mars (i.e. polar caps, frost, VFFs, etc) is both spatially widespread and diverse in form (**Table 3.2**). Questions remain regarding the depth, physical properties, and distribution of ground-ice in the subsurface on Mars. They also illustrate the multi-decadal evolution of our understanding, from early imaging of the polar caps to the current integration of radar, thermal, neutron, and high-resolution imaging data to understand the subsurface. However, until today, uncertainties remain regarding the depth, purity, volume, and long-term stability of this ice.

3.2.2 Obliquity-Driven Ice Deposition on Mars: *How does ice form on Mars?*

The current hypothesis of the primary driver of mid-latitude ice accumulation on Mars is the climate condition(s) associated with periods of high *mean* obliquity, during which polar ice becomes unstable and is redistributed as snow or frost towards lower latitudes (**Figure 3.12; Figure 3.13**). Unlike Earth's relatively stable axial tilt ($\pm 1^\circ$; Berger, 1988), Mars experiences dramatic variations in obliquity (tilt ranging from $\sim 15^\circ$ to over 35°) over 100,000- to million-year timescales due to the absence of a large stabilizing moon (and gravitational perturbations from Jupiter; Laskar et al., 2004). As shown in **Figure 3.13A**, Mars' obliquity is variable, meaning that beyond ~ 10 million years, the precise orbital history cannot be predicted, introducing uncertainty in long-term paleoclimate reconstructions. **Figure 3.13B** illustrates changes in Mars' orbital eccentricity, which modulates the intensity of seasonal insolation. However, its influence is secondary to obliquity in driving ice redistribution. However, over the last 10 million years, the average obliquity has remained relatively high, favoring recurring conditions that support ice accumulation in the mid-latitudes (Watanabe et al., 2024). During periods of high obliquity, polar regions receive increased solar radiation, causing net sublimation of the polar ice caps. Water vapour released into the atmosphere is then transported toward lower latitudes, where it condenses and accumulates as snow or frost particularly in shaded slopes, crater interiors, and high-relief terrains in mid-latitudes (Emmett et al., 2020; Madeleine et al., 2009; Vos et al., 2023).

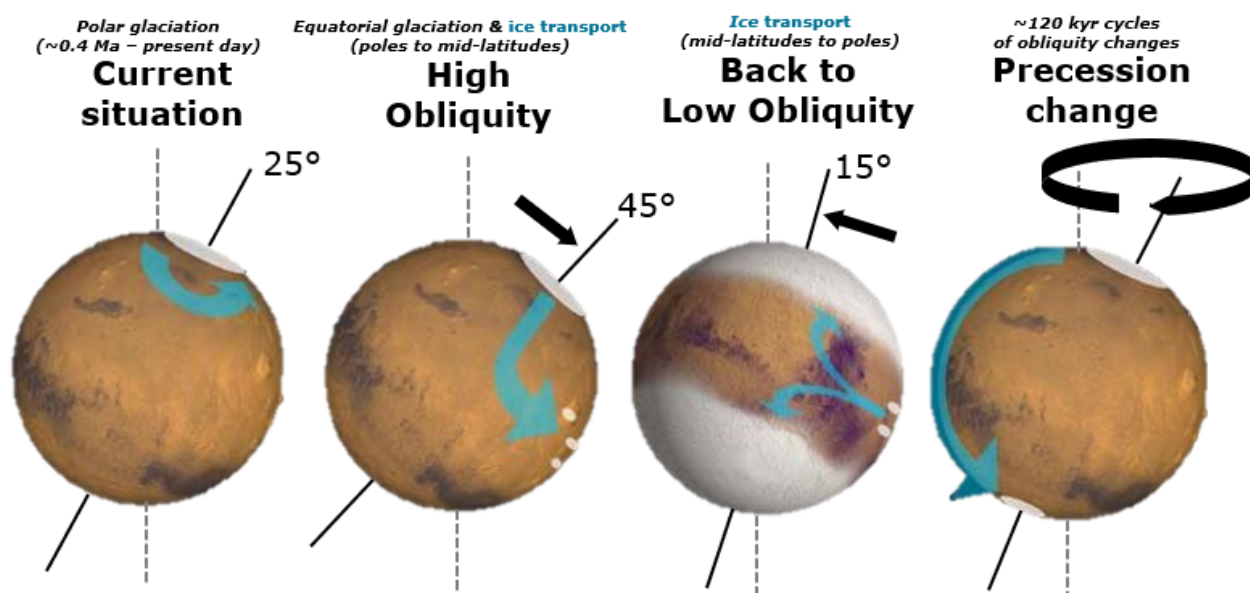


Figure 3.12. A summary of the recent (*last millions of years*) changes in Mars' orbital parameters and their consequences on the mobilization of water across the planet. In the first cartoon on the left, the blue arrow is reoriented to the north pole to indicate the seasonal sublimation and return of water to the permanent cap. All other cases with high, low obliquity and precession changes generate a permanent extraction of water from the cap towards locations indicated by the arrows. Redrawn by Chimira Andres, adapted from (Montmessin, 2007).

While individual spikes of high obliquity drive temporary ice redistribution, successive high mean obliquity cycle enable the long-term ice preservation and landscape modification of the mid-latitudes. These sustained high mean obliquities are thought to have resulted in the multiple episodes of glaciation evidenced by the deposition of ice-rich mantles and glacial landforms in the mid-latitudes, particularly during the Amazonian era (~2.1 Ga - present; **Figure 3.15**). These deposits, once formed, are preserved during lower-obliquity periods due to sublimation lag deposits (i.e. dust and debris that accumulate on the ice surface and act as an insulating layer (Bramson et al., 2017). This climate-driven mechanism explains the broad spatial distribution of Martian glaciers and mantled appearance of the terrain at $>40^\circ$ latitude. These features, typically found in regions of high relief and/or poleward facing slopes, reflect repeated accumulation and partial preservation of ice beneath thin lag deposits (**Figure 3.15**). The mantled terrains and “smooth”, flow-like morphologies further support the presence of buried, ice-rich materials. Preservation of ice is then enhanced during low-obliquity intervals when atmospheric water vapour decreases and polar sublimation slows, slowing down snow/frost redistribution, thereby, allowing debris layers to accumulate and shield the ice beneath. As a result, these mid-latitude ice deposits act as both paleoclimate archives and potential water resources for future exploration. Their distribution, longevity, and morphology offer key constraints on Mars’ recent climate history and its dynamic hydrological system.

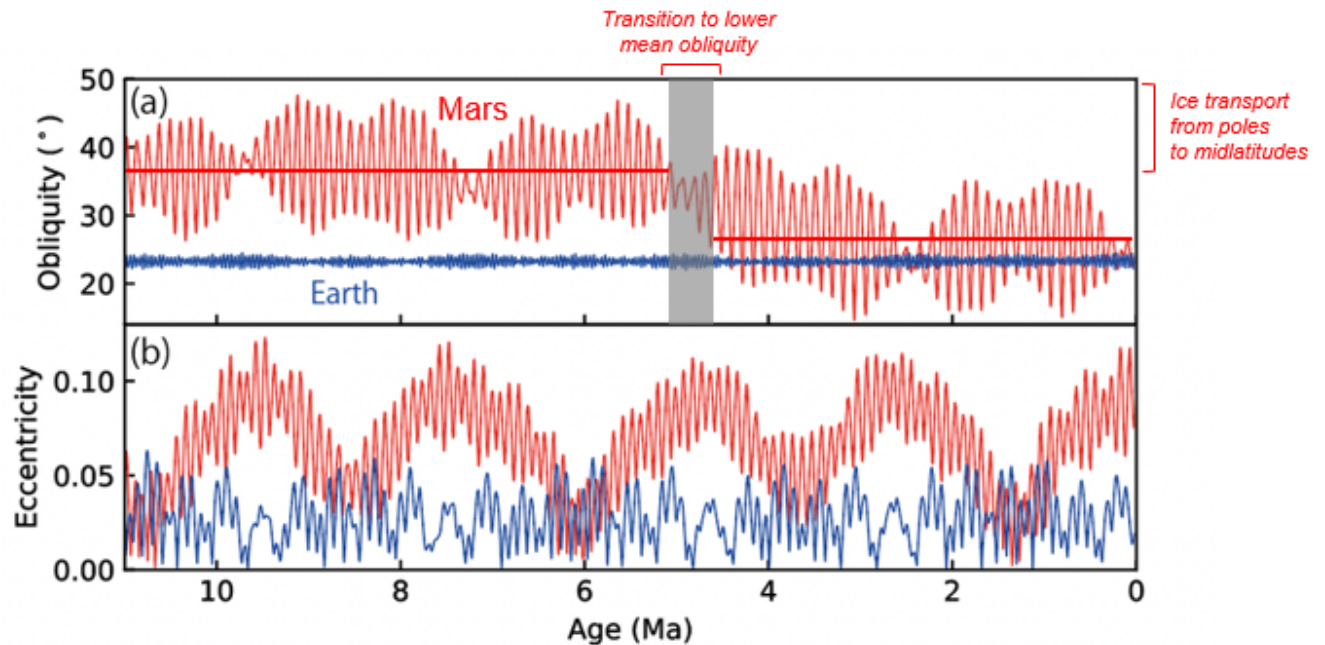


Figure 3.13. Variations of obliquity (a) and eccentricity (b) of Earth in comparison to Mars in the recent past (blue and red lines, respectively). Mars shows a more chaotic obliquity compared to Earth’s generally constant cycles. Modified from (Watanabe et al., 2024)

3.2.2.1 Mars Glaciation Eras

Most evidence for glaciation on Mars is associated with the Amazonian period, the current geological epoch (**Figure 3.14**). Amazonian glaciation is thought to have occurred in multiple, discrete episodes, each linked to peaks in mean obliquity (Buhler and Piqueux, 2021; Head et al., 2003; Laskar et al., 2004; Levrard et al., 2007). While high-obliquity peaks fluctuate every ~ 100 kyr (as shown in **Figure 3.13A**), only certain extended periods of sustained high mean obliquity and where climate conditions permit ice accumulation and preservation (i.e. cold surface temperatures, low sublimation rates, and sufficient atmospheric moisture) are hypothesized to result in widespread mid-latitude glaciation(s). Shorter or more variable high-obliquity episodes may not sustain the conditions required for extensive or long-lived ice deposits. Major glacial episodes are estimated to be around ~ 2.1 Ga, 1.4 Ga, 640 Ma, and 370 Ma, based on landform degradation states and crater counting. Crater counting, more formally known as crater size frequency distribution (CSFD) analysis, is a technique used to estimate surface ages by measuring the number and size distribution of impact craters (Michael and Neukum, 2010). However, on icy or volatile-rich surfaces, impact craters are easily degraded or erased by sublimation or resurfacing processes, making CSFD-derived ages approximate and potentially underestimated:

- *~ 2.1 Ga (Early Amazonian)*: Marks the earliest hypothesized episode of Amazonian glaciation. Evidence is sparse and heavily degraded, suggesting significant landscape modification since its formation (Levy et al., 2007).
- *~ 1.4 Ga (Middle Amazonian)*: A possible widespread glacial phase, inferred from moderately preserved landforms in the mid-latitudes (Head et al., 2006).
- *~ 640 Ma (Late Amazonian)*: More distinct glacial morphologies are observed from this period, including better-defined debris-covered glaciers and potential accumulation zones (Hepburn et al., 2020b).
- *~ 450 Ma (Very Late Amazonian)*: Represents the most recent and best-preserved glaciation episode. Many intact VFFs likely formed or were reactivated during this time, supported by clear SHARAD radar reflections and minimal surface degradation (Baker and Carter, 2019; Berman et al., 2015; Sinha and Vijayan, 2017).

During these episodes, large-scale atmospheric water redistribution led to the accumulation of glacial ice across the mid-latitudes, forming features like VFFs that persist today. Observational

studies have confirmed widespread occurrence and diversity of VFFs across various regions (Table 3.3). Initially, Squyres (1978) and Lucchitta (1984) identified extensive VFFs in the 30°–50°N range with estimated ages of ~10–100 Ma. More refined crater counting in Deuteronilus and Protonilus Mensae estimates VFF formation between ~10–20 Ma (Mangold, 2003), while in eastern Hellas, Pierce and Crown (2003) dated debris aprons to ~20 Ma. Younger features, such as polygonized and scaly GLFs in Hellas Basin, are thought to be no older than ~10 Ma (Hubbard et al., 2011). While older, Hesperian or even Noachian glaciation may have occurred (Figure 3.14), evidence is fragmentary and less well preserved (Dickson et al., 2008). Most ancient glacial landforms (older than Amazonian ages) outside of polar regions have likely been buried, reworked, or eroded. In contrast, Amazonian glaciers, being younger and better preserved, retain clear morphological signatures and basal radar reflections consistent with surviving ice bodies (Petersen et al., 2018a, 2018b; Plaut et al., 2009).

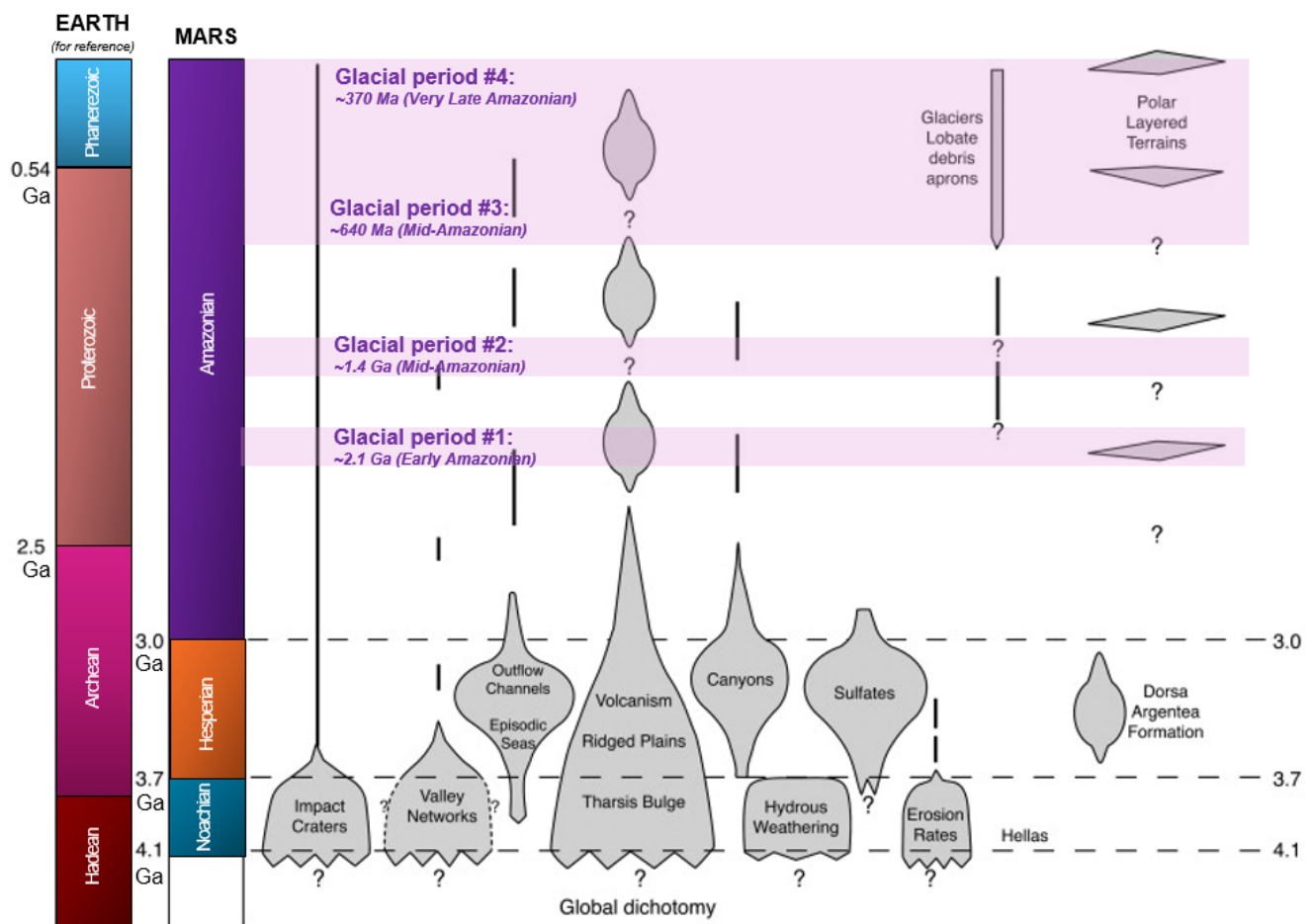


Figure 3.14. Geological activity on Mars over time, showing the relative importance of impact cratering, volcanism, landform development, and weathering, alongside major Martian time periods and their comparison to Earth's geologic eras, colours correspond to Figure 2.1 (Carr and Head, 2010).

Carr and Head (2010) emphasize that while the Amazonian is marked by reduced volcanic and fluvial activity, it is uniquely characterized by the formation and preservation of glacial features. These include VFFs especially on the flanks of large volcanoes like Olympus Mons as well as more broadly distributed across mid-latitudes between $\sim 30\text{--}55^\circ$ N/S. They are interpreted as the result of ice accumulation and downslope flow under obliquity-driven climate regimes. During periods of high average obliquity ($>35^\circ$), increased polar insolation leads to the redistribution of volatiles to lower latitudes (**Figure 3.12**), where ice becomes stable and accumulates (see also Section 3.2.2.1; Fastook et al., 2011; Laskar et al., 2004; Madeleine et al., 2009; references within Smith, 2022). This climate-driven cycling results in glacial advance and retreat phases preserved in mid-latitude terrains SHARAD data confirms that many of these glacial-like features retain near-pure ice beneath a protective debris cover, reinforcing their significance as cryospheric archives (Baker and Carter, 2019; Gallagher et al., 2021; Holt et al., 2008; Petersen et al., 2018b; Plaut et al., 2009).

3.2.3 Viscous Flow Features (VFFs): *Is There Active Glaciation on Mars?*

Viscous Flow Features (VFFs) is a descriptive term introduced by Milliken et al. (2003) to encompass a range of ice-related landforms on Mars that exhibit evidence of slow, gravity-driven deformation resembling glacial flow on Earth (**Figure 3.15**). These features typically exhibit concave-up headwalls, lineated or ridged surface textures, and lobate termini, suggesting viscous flow similar to terrestrial rock glaciers or debris-covered alpine glaciers. VFFs commonly originate from steep slopes and proposed that their morphology reflects viscous creep of ice-rich material under past climatic conditions favorable for ice stability (Dickson et al., 2008; Head et al., 2006; Hubbard et al., 2011).

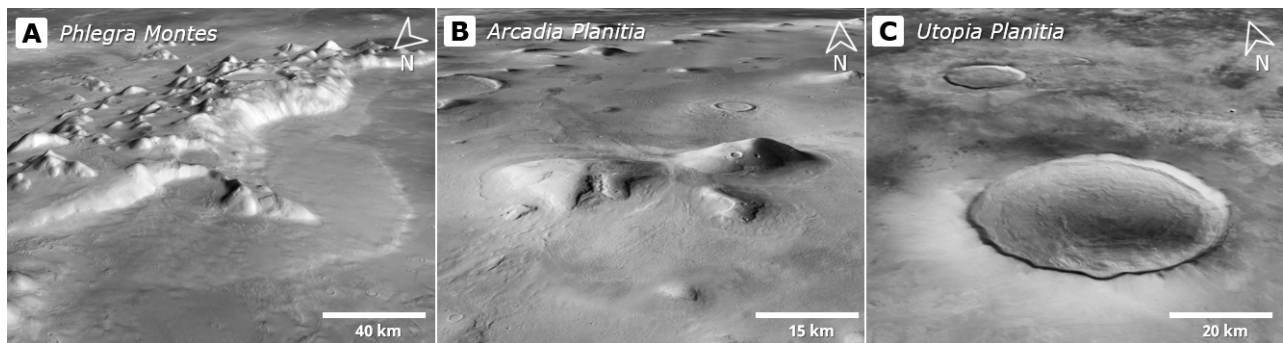


Figure 3.15. Examples of VFF morphologies in mid-latitude regions (also the three proposed landing sites of the Mars Life Explorer mission, expanded on in Section 3.2.6). MOLA-HRSC topography overlain by CTX image mosaic with a vertical exaggeration of 2. (A) An LDA at Phlegra Montes adjacent to the main mountain ridge. (B)

An LDA in Arcadia Planitia with smoothened boundaries due to the presence of the LDM. (C) A CCF in Utopia Planitia with concentric patterns within the crater.

The term, *VFF*, is an umbrella term capturing the various glacier-types on Mars such as lobate debris aprons (LDAs), lineated valley fills (LVFs), concentric crater fills (CCFs), glacier-like forms (GLFs), and composite systems (COMPs), which are combinations of the aforementioned VFFs (**Table 3.3**). These glacial landforms are concentrated primarily between 25-60° latitudes in both hemispheres and suggest long-term accumulation of glacial material (**Table 3.3**; **Figure 3.15 and 3.16**; (Levy et al., 2014; Souness and Hubbard, 2012). *LDAs* are tongue-shaped, ice-rich debris deposits that extend outward from the base of slopes, such as mesas, escarpments or crater walls (**Figure 3.15A**). *LVFs* are flow-like features that fill entire valleys, often with surface lineations oriented along the valley axis. *CCFs* appear as domed or concentric ridged deposits within craters, suggesting inward ice flow (**Figure 3.15C**). *GLFs* are smaller, isolated lobate forms found on slopes or within valleys that resemble alpine glaciers on Earth. *COMPs* represent systems where two or more VFF types interact or co-occur (e.g. an LDA feeding into an LVF). This type of glacier is most common in mountainous regions where there is complex topographic relief. Radar data and CSFD analysis or crater counting confirm the preservation of buried ice in VFFs, some of which remain morphologically pristine (Holt et al., 2008; Hepburn et al., 2020).

Table 3.3. Summary of VFFs and their contextual history inspiring their nomenclature from observation.

Feature	First Defined By	Observation
Lobate Debris Aprons (LDAs)	(Lucchitta, 1984)	Lobate features at the base of slopes or escarpments
Lineated Valley Fill (LVF)	(Lucchitta, 1984; Squyres, 1978)	Valleys filled with parallel or curvilinear ridges aligned with valley axes
Concentric Crater Fill (CCF)	(Levy et al., 2009)	Crater interiors with concentric ridges and smooth surface textures
Viscous Flow Features (VFFs)	(Milliken et al., 2003)	Umbrella term for LDAs, LVF, and CCF based on shared “flow-like” morphology
Glacier-Like Forms (GLFs)	(Souness et al., 2012)	Smaller-scale “flow” features in mid-latitude alcoves, often with lobe-shaped termini and parallel ridges
Composite/ Compound Systems (COMP)	(Gallagher et al., 2021)	Landforms showing superposition or interaction between multiple VFF units (e.g., CCF + LDA)

Mars currently lacks precipitation-fed glacial systems like those on Earth. However, glacial landforms such as viscous flow features (VFFs), including lobate debris aprons and glacier-like forms, preserve morphological characteristics such as sharp margins, flow-like textures, and convex profiles that are indicative of past glacial movement during the Amazonian period; **Figure 3.14**; **Table 3.4**).

Radar observations from the SHARAD instrument have revealed reflectors at the glacier base and dielectric properties consistent with buried, decameter-thick, high-purity ice within these landforms (Plaut et al., 2009; Holt et al., 2008). While this confirms ice preservation, it does not demonstrate current activity. Despite these observations, there is no direct evidence that glacial landforms on Mars are currently active. Instead, present-day conditions appear to favor sublimation of existing ice rather than net accumulation (Levy et al., 2014). Surface features such as fresh shallow valleys (FSVs; Hobley et al., 2014; Wilson et al., 2016) or esker-like ridges (Butcher et al., 2016; Gallagher and Balme, 2015) are interpreted as relics of past deformation, not signs of active movement. Thus, Martian glaciers are considered relict as they retain form and internal ice but lack dynamic behavior under current climatic conditions or could be too slow to observe with the timescale of current Mars missions (Anderson et al., 2018; Mahaney et al., 2007).

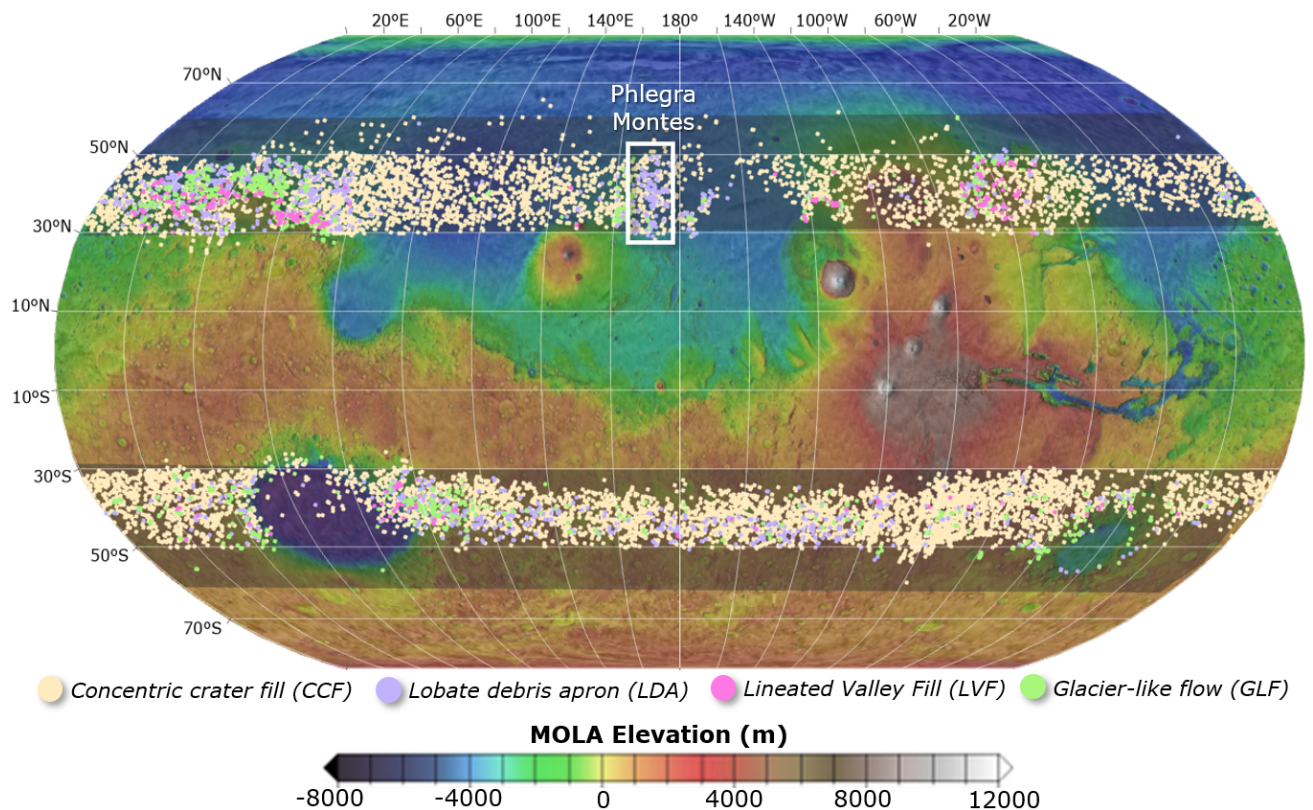


Figure 3.16. Distribution of viscous flow features (VFFs) on Mars. Map is centered on Phlegra Montes. Data sources: (Driver et al., 2024; Levy et al., 2014)

VFF morphology, stratigraphy, and regional distribution all point to a history of climate-driven ice accumulation and partial preservation, rather than active glaciation under current Martian conditions. The identification and analysis of VFFs has significantly advanced our understanding

of Mars' cryospheric history, revealing that mid-latitude ice is spatially extensive. Their preservation under insulating debris suggests that these features may house accessible water-ice for future exploration and serve as valuable analogues for studying ice-related processes on Earth and other icy bodies.

3.2.4 Latitude Dependent Mantle (LDM)

The Latitude Dependent Mantle (LDM) is an ice-rich deposit that drapes the Martian surface poleward of $\sim 30^\circ$ latitude in both hemispheres, with best preservation beyond $\sim 45^\circ$ – 50° (**Table 3.4**; Head et al., 2003; Mustard et al., 2001). The LDM is thought to be formed through the episodic deposition of dust and snow during obliquity excursions, particularly when Mars' axial tilt exceeded ~ 35 – 45° , rather than during prolonged periods of high mean obliquity (Forget et al., 2006; Milliken et al., 2003). A study by Berman et al. (2009) also hypothesizes that some VFFs may represent former LDM deposits that flowed under topographic confinement (i.e. lobate flows within crater walls with arcuate ridges). Its smooth texture, soft appearance, and pervasive coverage suggest airfall origin, while surface features such as thermal contraction polygons along with neutron data support the interpretation that it is comprised of near-surface ice (Boynton et al., 2002; Hiesinger and Head, 2000; J. Levy et al., 2009; Seibert and Kargel, 2001).

HiRISE and recent SHARAD observations (Xu et al., 2025) reveal that the LDM ranges from a few meters to over 100 meters in thickness, exposing relatively clean ice beneath a thin debris lag in steep scarps (i.e. $>50^\circ$ N) and crater walls (Baker and Carter, 2023; Conway and Balme, 2014; Dundas et al., 2023; Khuller and Christensen, 2021). Its degradation is attributed to sublimation over time (Lefort et al., 2009; Kress and Head, 2014). It lacks any evidence of flow or deformation making it distinct from VFFs. Age estimates derived from CSFD analysis range from ~ 20 million years in older, high-latitude deposits to as young as ~ 0.35 – 2.1 million years (Head et al., 2003; Kostama et al., 2006; Schon et al., 2012; Willmes et al., 2012). The LDM is estimated to cover 23% of the surface of Mars and to store $\sim 10^4$ – 10^5 km³ water-ice volume (Kreslavsky and Head, 2002), making it a key potential resource for future exploration specifically in mid- to high-latitude regions where it is likely to be better preserved than degraded, ice-poor deposits observed at lower latitudes (Clifford and Hillel, 1983; Vos et al., 2019).

Table 3.4. Literature on Mars that studied mid-latitude VFFs and LDM. This is limited to studies that analysed the ages for the LDM and VFFs.

Author	Latitude range of Study	Feature & Textures Observed	Estimated Age
(Squyres, 1978)	30°-50°N	“Fretted terrain” Lobate debris aprons, lineated valley fill	~100 My (Late Amazonian)
(Lucchitta, 1984)	30°-50°N	Glacial landforms, evidence of debris-covered glaciers	~10-100 My (Late Amazonian)
(Mustard et al., 2001)	~30°–60°N	TES-based thermal inertia of mantling deposits, consistent with ice-rich material	Amazonian (relative stratigraphy)
(Kreslavsky and Head, 2002)	30°–60°N	Mapping of mantle smoothness; latitude-dependent patterns	-
(Mangold, 2003)	35°–50°N (Deuteronilus / Protonilus Mensae)	LDAs, geomorphic analysis in “fretted” terrain from ice sublimation	~10-20 My (Late Amazonian)
(Pierce and Crown, 2003)	Eastern Hellas Region	Morphology and topography of debris aprons;	~20 My (Late Amazonian)
Head et al., 2003	30°–60°N	LDM: smooth, layered, ice-rich mantle; obliquity-driven deposition and degradation	~0.4–20 My (Mid- to Late Amazonian)
(Milliken et al., 2003)	30°-60°N	Layered and smooth textures of ice-rich LDM; Mantle deposition and glacial flow processes	~10 My (Mid- to Late Amazonian)
(Forget et al., 2006)	30°-50°N 30°-50°S	Water-ice rich mantle, climatic modeling	~1-10 My (Mid- to Late Amazonian)
(J. Levy et al., 2009)	40-45°N (Utopia Planitia)	CCF: “brain terrain” and polygonal patterned ground	minimum age of ~1.5 My (best-fit ages span 1.3 to 1.7 My)
(Kadish et al., 2009)	30°–60°N	Mapping of pedestal craters suggesting widespread ice-rich mantle	Crater retention age: ~10–100 My
(Hubbard et al., 2011)	38.15°S, 113.16°E (Hellas Basin)	GLF: Scaly terrain, Polygonized terrain, Linear terrain, Mound-and-tail terrain, Incised-headwall terrain, Rectilinear-ridge terrain	not older than ~10 Ma
(Schon and Head, 2012)	30°–60°N	Polygonally fractured mantle textures, obliquity-linked cycles	~2–10 My (Late Amazonian)
(Conway and Balme, 2014)	45° and 33° N and S (excl. Hellas Basin, Argyre Basin, and Thaumasia regions)	Pole facing slopes, gullies incising mantle deposits	0.35 Ma to ~2.1 Ma
(Kress and Head, 2014)	30°-60°N	Degraded textures, remnant patches of ice, Transition of LDM through sublimation and degradation	~1-10 My (Late Amazonian)
(Berman et al., 2015)	40°-50°N	Lineations, troughs, and ridges on VFFs; Flow behavior of ice and surface deformation	~10 My (Mid- to Late Amazonian)
(Baker and Carter, 2019)	Mid-latitudes	Supraglacial debris: sources, thickness, stratigraphy	~1-2 My (Late Amazonian)

3.2.5 Periglacial Landforms on Mars

The Martian mid-latitudes host a range of landforms that closely resemble periglacial features found in Earth's Arctic and subarctic environments. These landforms offer insight into past and potentially ongoing cryospheric processes on Mars, particularly those linked to thermal contraction, sublimation, and degradation of ground ice (**Figure 3.17**; Mangold, 2003; Seibert and Kargel, 2001). Here, we mention only those landforms that are relevant to understand the content of this thesis (i.e. polygonal terrain, scalloped depressions, and brain terrain), as a complete review would be beyond the scope of this work.

Polygonal terrain is widespread poleward of $\sim 45^\circ$ and is especially common within the Latitude Dependent Mantle (LDM). It is interpreted as the result of thermal contraction cracking in ice-rich regolith, forming seasonal or multi-annual polygons as cracks fill with dust or frost (Levy et al., 2009; Mellon et al., 2004; Philippe et al., 2022; Ulrich et al., 2011). These features range from 5 to 30 meters in width and are clearly visible in HiRISE and CTX imagery (**Figure 3.17A**). *Scalloped depressions* are crescent-shaped features with steep equator-facing walls and gentler pole-facing slopes. They are interpreted as sublimation thermokarst features formed by the loss of excess ice, particularly in pole-facing slopes of ice-rich deposits (**Figure 3.17B**; Costard and Kargel, 1995; Lefort et al., 2010). These landforms are especially common in Utopia Planitia, Malea Planum and northern Phlegra Montes. They will be discussed in Chapter 6 as one of the significant surface textures present on the LDM. Their morphology and spatial distribution are analogous to thermokarst depressions in terrestrial permafrost regions such as northern Canada and Siberia (Morgenstern et al., 2007; Murton, 1996; Ulrich et al., 2010). *Brain terrain* refers to labyrinthine, ridge-and-groove surface patterns interpreted as the result of ice sublimation within debris-covered ice deposits (**Figure 3.17C**; Levy et al., 2009). These features commonly occur alongside degraded glacial landforms and the LDM. Analogous features in Earth's high-latitude permafrost regions form where differential thawing and ice loss deform the surface over time (Bina and Osinski, 2021; Hibbard et al., 2021).

Collectively, these Martian periglacial-like features provide geomorphic evidence for relatively recent and climate-sensitive cryogenic processes. Their distribution supports the persistence of near-surface ground ice and reinforces the interpretation that Mars' mid-latitude landscapes remain strongly shaped by cryospheric activity during the Amazonian.

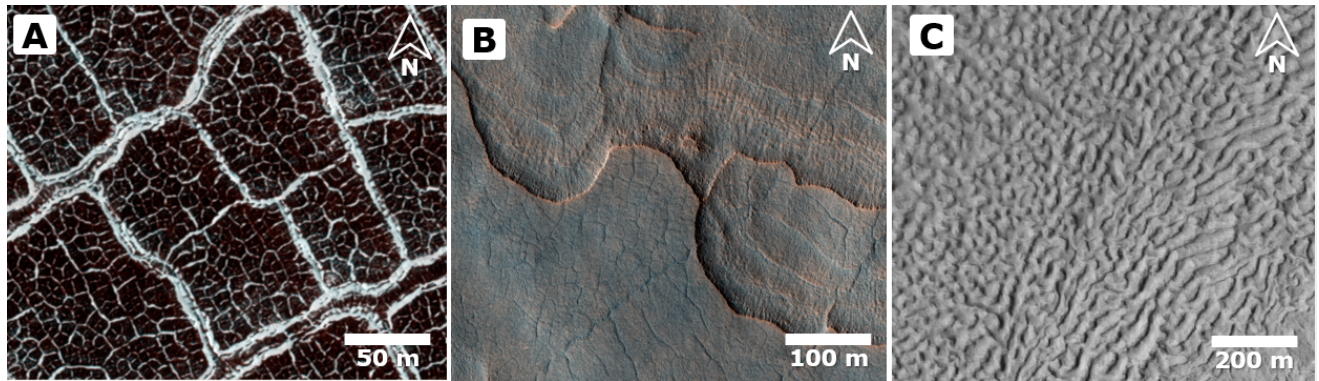


Figure 3.17. Periglacial analogue terrain on Mars illustrating the geomorphic diversity. **(A)** HiRISE image of polygonal patterned ground (Image ID: ESP_016641_2500). **(B)** HiRISE image of scalloped polygonal depressions interpreted as sublimation features in permafrost terrain (Image ID: PSP_002162_2260). **(C)** HiRISE image of brain terrain on the surface of a VFF in Deuteronilus Mensae (Image ID: ESP_045507_2200).

3.2.6 Mid-latitude ice

As Mars' mid-latitudes house numerous VFFs, interpreted as relics of cold-based glaciers, SHARAD radar data have confirmed that many contain massive, subsurface ice (i.e. <400 m in mean ice thickness; Andres and Smith *in review*, 2025) protected by debris or regolith, typically a few meters thick (Baker and Carter, 2019; Holt et al., 2008; Karlsson et al., 2015; Plaut et al., 2009). Efforts to estimate the volume of ground ice in Mars' mid-latitudes often aggregate both LDM and VFFs, as they frequently co-occur (see Sections 3.2.3 and 3.2.4). This complicates attempts to quantify the individual contributions of each unit to the overall mid-latitude ice reservoir.

Recent efforts by the Subsurface Water Ice Mapping (SWIM) project have significantly advanced our understanding of near-surface ice on Mars (Morgan et al., 2025, 2021). Using a data integration approach that combines multiple remote sensing datasets (i.e. neutron spectroscopy, thermal inertia, radar, and geomorphic indicators), the SWIM team has developed probabilistic maps of shallow ice occurrence, particularly in the mid-latitudes (**Figure 3.18**). Each dataset's result was translated into a value from -1 (inconsistent with ice) to +1 (fully consistent with ice), with depth-specific weightings. For example, in the 0–1 m range, thermal inertia and neutron data were emphasized. Thermal inertia, derived from THEMIS orbital infrared measurements, reflects how quickly a surface heats and cools (i.e. low thermal inertia typically indicates fine, porous, or ice-rich material; Christensen et al., 2004). On the other hand, Neutron spectroscopy data from the Mars Odyssey Neutron Spectrometer (MONS), detects hydrogen abundance (a proxy for water) by measuring the flux of neutrons emitted from the surface after being bombarded by cosmic rays, providing strong evidence for near-surface

ice (Boynton et al., 2004; Feldman et al., 2002). At greater depths, radar-based dielectric analysis (i.e. SHARAD or MARSIS) and geomorphic indicators of ice-related landforms (e.g., VFFs and polygons) were weighted more heavily (methods detailed in Morgan et al., 2025, 2021).

The final maps were produced by combining these weighted scores at a resolution of ~ 3 km / pixel. These maps not only predict the widespread nature of excess ice beneath thin protective covers but also identify regions with the highest potential for accessible ice, critical for mission planning. By integrating these datasets and applying consistent analytical criteria across them, the SWIM initiative provides one of the most robust and comprehensive assessments of Martian subsurface ice to date. Their results support the hypothesis that some mid-latitude regions host stable, decameter-thick ice deposits just below the surface, supporting the case for both scientific and future human exploration.

Despite these advances, several uncertainties limit the precision and interpretability of SWIM's ice consistency maps (**Figure 3.18**). First, the vertical resolution of contributing datasets varies significantly. While the 0–1 m range (**Figure 3.18A**) benefits from high-resolution data such as thermal inertia and neutron spectroscopy, deeper intervals (1–5 m and >5 m; **Figure 3.18B** and **3.18C**) depend more heavily on radar-based dielectric analysis, which has coarser vertical resolution and limited ability to resolve layering or differentiate between pure and debris-rich ice. Second, integrating heterogeneous datasets, each with different spatial resolutions, coverage, mapping biases (i.e. mapped geomorphology databases) and sensitivities, can introduce inconsistencies and bias, especially in regions with sparse or noisy data. Moreover, while high radar reflectivity or hydrogen abundance often suggests ice presence, the actual purity, porosity, and extractability of that ice remain unknown. These uncertainties are particularly relevant for assessing in-situ resource potential, where debris content, depth of burial, and thermal stability are critical. As such, while the SWIM maps provide a valuable, probabilistic framework for identifying promising ice-rich regions, the exact distribution, composition, and accessibility of Martian subsurface ice remain open questions, highlighting the need for targeted future missions with higher-resolution instruments and in situ verification capabilities (Putzig et al., 2023).

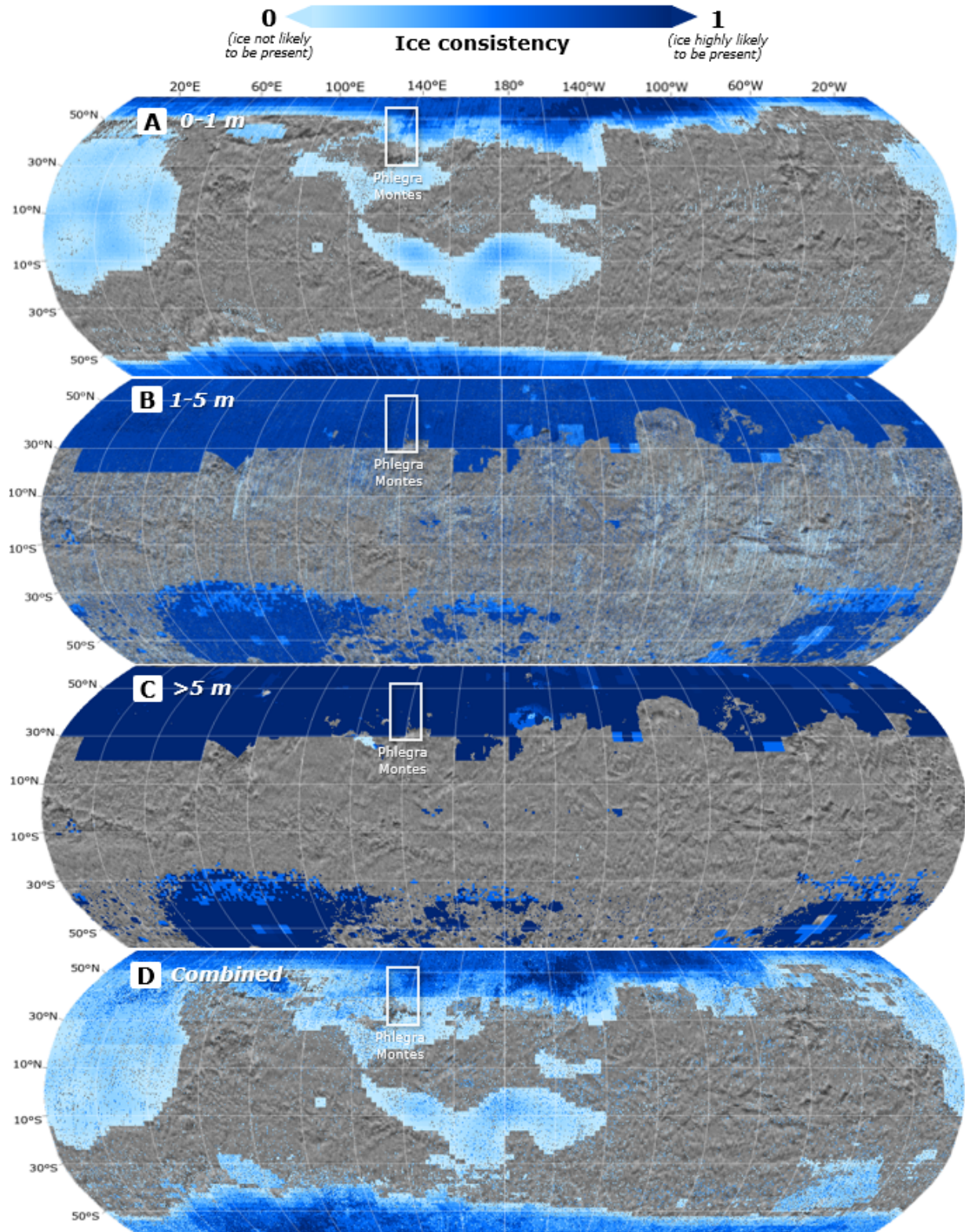


Figure 3.18. Ice consistency maps from the SWIM 2.0 project showing the likelihood of subsurface ice on Mars at (A) 0–1 m, (B) 1–5 m, (C) >5 m depths, and (D) combined. Values range from 0 (unlikely) to 1 (highly likely). Phlegra Montes (outlined) shows consistently high ice potential across depths. Image credit: SWIM 2.0, PSI/NASA.

3.2.7 In-situ resource utilization (ISRU)

In-situ resource utilization (ISRU) is central to enabling sustained human presence on Mars, with shallow, accessible water-ice in the mid- to high-latitudes being key targets for life support systems and fuel production. Ideal ISRU sites require relatively pure, near-surface ice with minimal overlying debris as it is far more efficient to extract and process than ice mixed with regolith, which requires additional energy and equipment to separate usable water; Sanders et al., 2015). The structural integrity of ice bodies (i.e. whether they are stratified, massive, or debris-thickness vs ice thickness) also affects the choice of extraction techniques and the mechanical feasibility of drilling or trenching operations. Additionally, locations closer to the equator (lower latitudes) are favored due to more moderate temperatures and increased solar insolation, which improve thermal efficiency and solar power availability for surface operations (Davis et al., 2022). The widespread occurrence of decameter-thick ice in Mars' mid-latitudes, documented by SHARAD and synthesized in SWIM's probabilistic maps, positions these regions as strong candidates for ISRU operations. However, our ability to further refine ice distribution, depth, and physical properties is limited by an aging fleet of orbital instruments, including HiRISE, SHARAD, MARSIS, and MONS, all of which are operating beyond their planned lifespans.

To address this, the proposed *International-Mars Ice Mapper (I-MIM)* mission would provide high-resolution, planet-wide mapping of near-surface ice (upper 1-5 m) using SAR as the main anchor payload (I-MIM MDT, 2022). This data would enable improved ISRU site selection, inform the design of resource extraction systems, and support mission safety through better terrain understanding. Additionally, I-MIM is positioned to act as a precursor to more targeted missions like *Mars Life Explorer (MLE)*, which aims to investigate the astrobiological potential of near-surface ice through deep drilling and biomarker detection in regions of high-habitability potential (i.e. mid- to high-latitudes; Phillips-Lander, 2024). Together, mission concepts such as I-MIM and MLE for example, represent a new phase of Mars exploration, one that considers Mars as an integrated system, where cryospheric processes, climate history, resource potential, and astrobiology intersect. Replacing and expanding our orbital capabilities is essential not only for advancing science but also for ensuring the viability of future human and robotic missions.

3.2.8 Does Ice on Mars melt?

The melting of ice on Mars remains an active area of research. Under present surface conditions, characterized by low atmospheric pressure (~6 mbar), subzero temperatures, and low humidity, liquid water is generally thermodynamically unstable and transitions directly from solid to vapour via sublimation (Hecht, 2002a, 2002b). As a result, stable surface melt is highly unlikely today (**Figure 3.19**). However, transient melting may occur under specific environmental conditions. Experimental and thermal modeling studies suggest that thin insulating layers of regolith or dust can trap heat and create localized thermal anomalies, enabling brief episodes of subsurface melting (Clow, 1987; Khuller and Clow, 2024; Zent and Quinn, 1995). Additionally, salt-rich environments can depress the freezing point of water, forming metastable brines capable of persisting transiently under Martian conditions (**Figure 3.19**; Chevrier and Rivera-Valentin, 2012; Martín-Torres et al., 2015).

In Mars' past, particularly during high-obliquity periods or early epochs when atmospheric pressure was higher, the potential for surface melting was greater (Mischna et al., 2003; Wordsworth, 2016). Meltwater generation may have occurred on equator-facing slopes, in crater interiors, or beneath glacial debris where insolation and topography favored warming (Fastook et al., 2012, 2008; Madeleine et al., 2014). Although widespread surface melting is unlikely today, localized melt events especially in buried ice or brine-rich regions, remain plausible.

Subsurface environments insulated by overlying debris can retain enough geothermal heat to induce basal melting, especially where brines depress the freezing point of water. For example, some studies of the south polar layered deposits (SPLD) suggest the presence of subglacial liquid water, potentially maintained by localized geothermal anomalies (Arnold et al., 2022; Orosei et al., 2018). Similarly, interpretations of esker-like features close to mid-latitude viscous flow features by Butcher et al., (2016) and Gallagher and Balme (2015) propose past episodes of basal melting, driven by elevated geothermal flux or pressure-induced melting beneath cold-based glaciers. These environments can have potentially significant implications for astrobiology and future exploration, as they could represent rare habitable niches where liquid water may still transiently exist, which remains an open question.

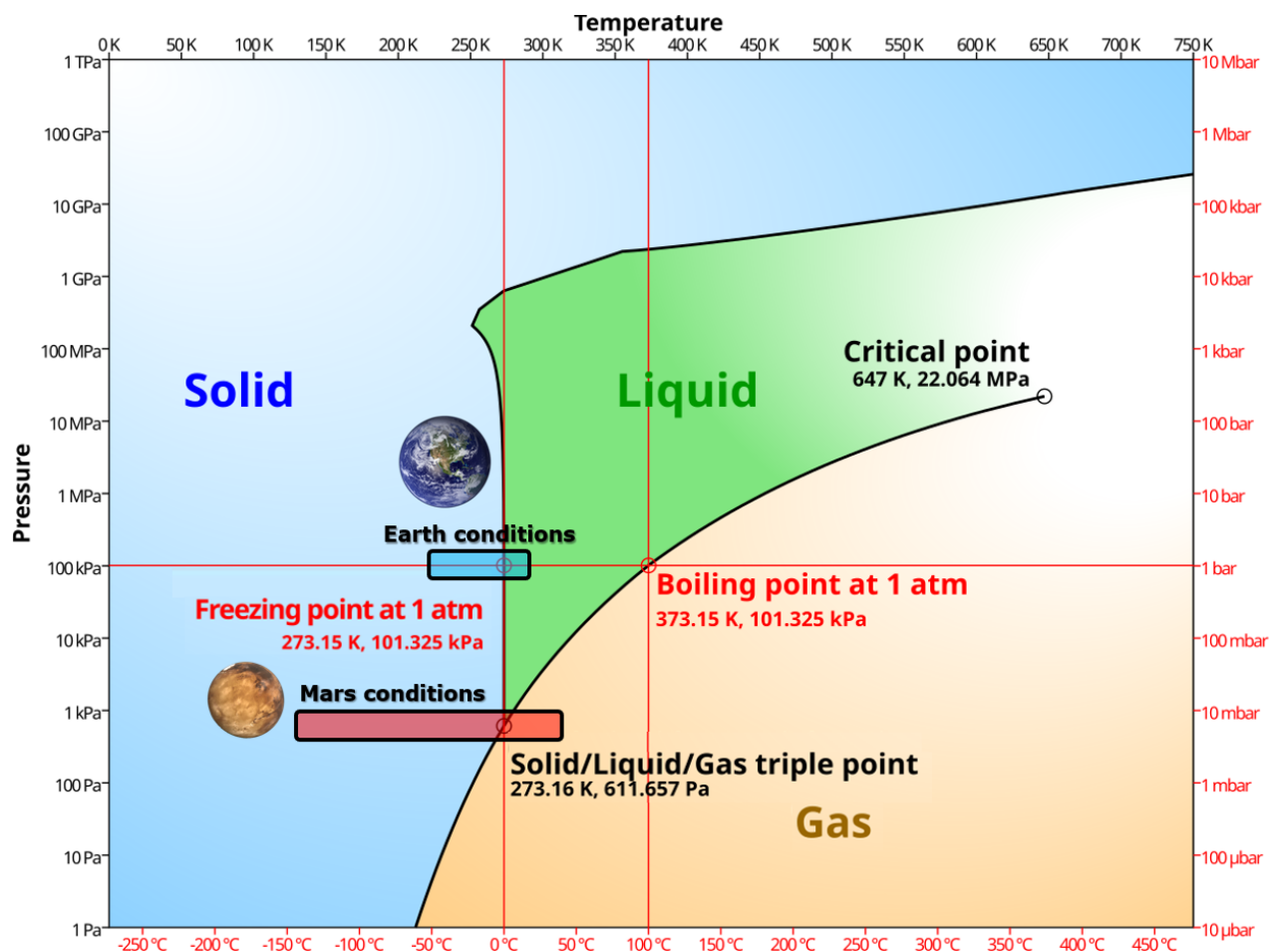


Figure 3.19. Phase diagram of water, showing the stability fields of ice, liquid, and vapour. Highlighted regions represent typical surface conditions for Earth (blue) and Mars (red). Earth's surface spans ~250–310 K and ~1 bar (10^5 Pa), allowing stable liquid water. Mars exhibits a wide surface temperature range (~140–310 K) but maintains low atmospheric pressure (~6 mbar), placing most surface conditions below the triple point of water (~273.16 K, 6.1 mbar). As a result, water-ice on Mars typically sublimates rather than melts. However, transient liquid water may occur in localized settings with peak insolation or salt-rich substrates (Hecht, 2002; Martínez and Renno, 2013; Piqueux et al., 2016; Ojha et al., 2015).

3.3 Open Questions

Despite significant progress in understanding the Martian cryosphere, fundamental gaps remain in how we detect, characterize, and interpret the planet's surface and subsurface water-ice reservoirs. This dissertation addresses several of these unresolved challenges by combining radar sounding, geomorphologic analysis, and terrestrial analogue studies to refine our understanding of Mars' water budget and its implications for climate and habitability. The following open questions frame the core scientific inquiries addressed in the subsequent chapters.

3.3.1 Is There Internal Structure and Ice Heterogeneity in Mars' Mid-Latitudes?

Accurately detecting and characterizing near-surface water-ice, particularly its purity and internal structure, remains a critical challenge. SHARAD have identified kilometer-thick ice deposits in many mid-latitude settings, yet distinguishing clean from debris-covered ice is complicated by the limited resolution and penetration depth of current instruments. Recent efforts like the SWIM project (Morgan et al., 2025) integrate multi-sensor datasets to generate probabilistic maps of near-surface ice stability, but uncertainty persists, especially in geomorphologically and stratigraphically complex regions like Phlegra Montes. Upcoming missions like the Mars Ice Mapper (I-MIM) can perhaps address this issue globally, but a deeper regional analysis using existing SHARAD data is needed. This thesis tackles this knowledge gap by quantifying ice volumes and estimating debris lag thickness in Phlegra Montes through radar analysis and new analytical modeling (**Objective #1; Sub-objective #1-2**).

3.3.2 How Do We Characterize the *Glacial-Periglacial Continuum* on Mars in Relation to the VFFs and LDM?

There is ongoing discussion about the extent to which mid-latitude ice-rich landforms on Mars reflect *glacial* vs *periglacial* processes (i.e. VFFs vs LDM). This distinction is not just semantic as it has implications for interpreting Mars' climatic history. Glacial processes (associated with VFFs) imply snow accumulation, flow, and possibly meltwater, while periglacial landforms (associated with the LDM) could form under dry, cold conditions through sublimation and thermal contraction. Some features exhibit morphologies and internal structures consistent with glacial flow, while others more closely resemble permafrost features on Earth. More specifically, *what is the tipping point or points between glacial and non-glacial ice on Mars?* **Objective #2** directly addresses this question on the glacial-periglacial continuum on Mars by examining the geomorphologic and latitudinal variation of VFFs and their relationship with the LDM.

In addition, the genetic relationship between the LDM and VFFs is an open discussion. *Are they distinct co-occurring features, or do they represent different stages in an evolutionary continuum driven by obliquity-driven climate change?* For example, VFFs may represent former LDM that flowed under topographic confinement and/or around unconfined massifs. Alternatively, the LDM may be a later mantling phase that draped over older glacial features. Understanding this

relationship is essential for reconstructing Amazonian climate history. This will be investigated by examining their spatial associations and stratigraphic overlaps within Phlegra Montes (**Sub-objective 2.3**).

3.3.3 Are Martian Glaciers More Like Rock Glaciers or Debris-Covered Glaciers?

A related issue or discussion point is the classification of VFFs within the category of debris-covered glacier and rock glacier terrestrial analogues. Rock glaciers typically form through creep of ice-rich colluvium and fed by snowpack input, which may better match Mars' cold, dry climate. Debris-covered glaciers, are glacier ice bodies mantled by thick supraglacial debris that insulates the underlying ice, slows ablation, and enables flow under relatively low accumulation and temperature conditions. This debris cover often preserves internal stratigraphy that records episodic snowfall or ice recharge (as mentioned in Section 3.1.2.3).

On Mars, distinguishing between these analogues remains challenging due to current limitations in radar resolution. Instruments like SHARAD may not detect internal layering thinner than several meters (i.e. <15m), meaning finer-scale structures could remain unresolved. To better determine whether VFFs resemble debris-covered glaciers with preserved accumulation history or more permafrost-like rock glaciers with limited recharge, higher resolution subsurface data are needed. This is explored in both radar-based volume analysis (**Objective #1**) and morphologic classification efforts (**Objective #2**) of VFFs.

3.3.4 How Effectively Can Terrestrial Analogue Studies Constrain the Detection and Classification of Ground Ice on Mars?

Understanding how ground ice is expressed and detected in terrestrial permafrost environments is essential for improving ice detection strategies on Mars. *How do different geophysical methods (i.e. GPR, CCR, and airborne radar) perform in characterizing shallow ice types, and what are their limitations? Can we reliably distinguish between excess ice, ice wedges, and buried glacial ice using surface-based and airborne techniques?* Insights from analogue sites like Tombstone Territorial Park are key to validating radar interpretations, informing Mars Ice Mapper mission planning, and refining our understanding of cryospheric processes on both Earth and Mars (**Objective #3**).

CHAPTER 4

4 Regional ice-depth and thickness in Phlegra Montes, Mars from radar characterization of glacial landsystems using SHARAD

4.1 Introduction

4.1.1 Mid-latitude Ice on Mars

Mars is a frozen world where water-ice is abundant above, at, and under the surface. Radar studies performed to date have substantiated the existence of shallow ground ice and have confirmed the existence of debris covered glaciers on Mars (Holt et al., 2008; Plaut et al., 2009; Berman et al., 2015; Petersen et al., 2018). Observations of transient ice exposures have led to speculation that large quantities of H₂O ice may be present in Mars' midlatitudes, preserved beneath a layer of dust or regolith and thus shielded from the atmosphere (Bramson et al., 2017). The mid-latitudes range from 30°-60° in both northern and southern hemispheres (**Figure 4.1**).

In Mars' mid-latitudes, viscous flow features (VFFs) are widely thought to be debris-covered glaciers (Levy et al., 2014; Milliken et al., 2003). The term VFF refers to a variety of glacier-types, including glacier-like forms (Brough et al., 2016, 2019; Driver et al., 2024; Hepburn et al., 2020b), lobate debris aprons (LDAs), lineated valley fills, and concentric crater fills (Levy et al., 2010; Milliken et al., 2003; Squyres, 1979). The presence and interpretation of these debris-covered glaciers have been supported with evidence from several geophysical (Holt et al., 2008; Perry et al., 2023; Petersen et al., 2018c; Plaut et al., 2009), spectral (Grau Galofre et al., 2022), geomorphologic (Hubbard et al., 2011; Levy et al., 2014), and climate modeling (Fastook et al., 2008; Madeleine et al., 2009) studies. The surfaces of VFFs on Mars exhibit complex, often curvilinear patterns that likely record histories of ice flow with implications to the paleogeography of the region (Levy et al., 2021; Morgan et al., 2009). Similar to Earth's glaciers, the surface patterns of VFFs may reflect the complexities of their internal subsurface structure (Butcher et al., 2023; Levy et al., 2021).

Among all VFF morphologies on Mars, LDAs are the largest reservoir of ice (Levy et al., 2014). LDAs appear as gently sloping (~5-20°), lobed landforms and over long timescales and

conditions, have deformed (and flowed) leading to a characteristic lobate shape (Karlsson et al., 2015; Morgan et al., 2009; Parsons and Holt, 2016; Squyres, 1979). Known LDAs have a debris cover on their surfaces, which can insulate and protect the underlying ice from sublimation (Aharonson and Schorghofer, 2006; Chevrier et al., 2007; Fastook et al., 2014; Schorghofer and Aharonson, 2005, 2005). However, in regions with varying topographic relief such as in Phlegra Montes, several LDAs are fed by higher elevation, up-valley lineated valley fills and can also be collocated with concentric crater fills (Gallagher et al., 2021). A combination of a variety of these glacier morphologies create ice-rich composite systems. In this study, composite systems are defined to be a combination of VFFs merging into a single landform.

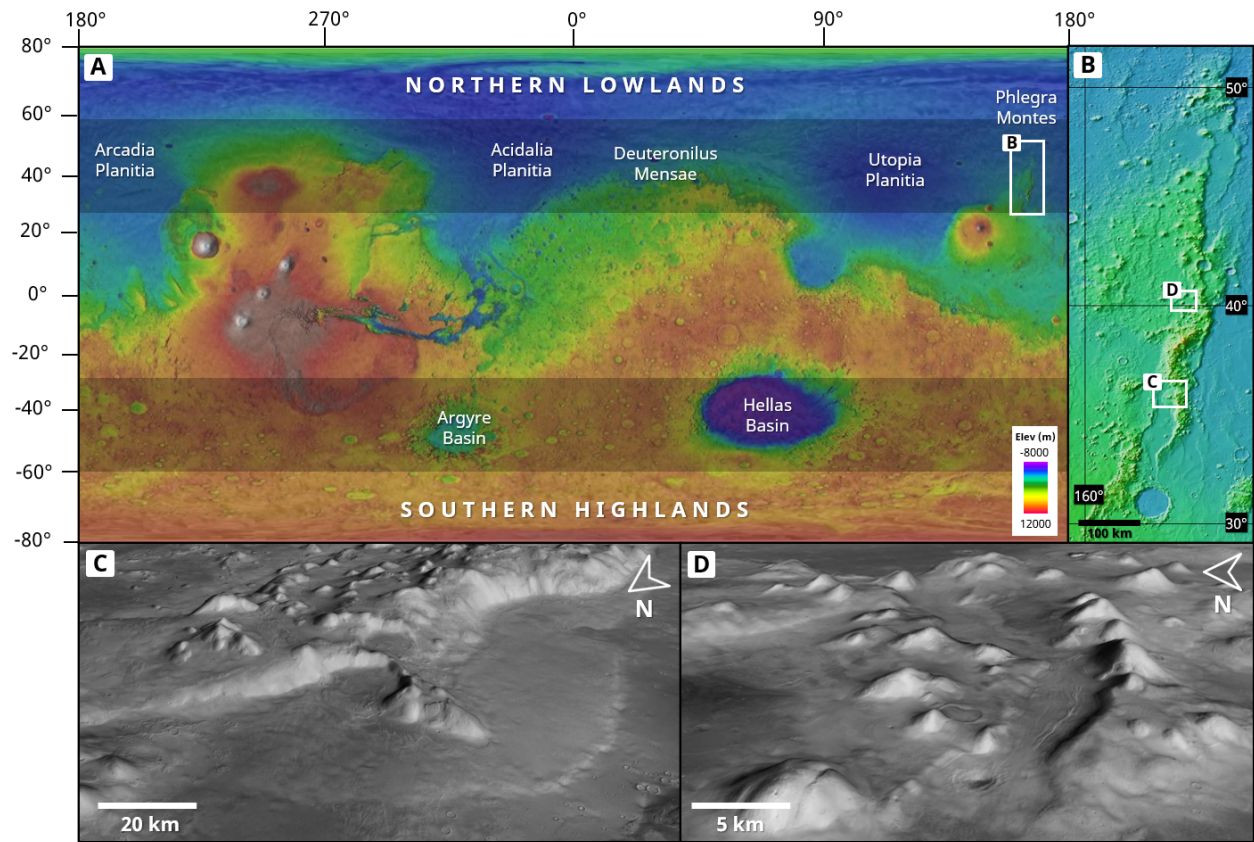


Figure 4.1. (A) Topography of Mars. Major features are identified, and latitude bands exhibiting VFFs are highlighted. The location of our study area denoted in the white boundary box. (B) MOLA Elevation map showing inset map of Phlegra Montes. (C) Oblique CTX overlay on MOLA-HRSC of a lobate debris apron (LDA) at 164° E, 35° N along the main ridge of Phlegra Montes (highlighted in Section 3.2.1). (D) Oblique CTX overlay on MOLA-HRSC of alienated valley fill (LVF) at 164° E, 40° N.

VFFs across Mars' mid latitudes are geologically young, having modeled surface ages on the order of millions to hundreds of millions of years (i.e. late Amazonian) (Baker and Head, 2015; Berman et al., 2021; Hepburn et al., 2020b). VFFs are expected and have been proposed to be a

target for future in-situ resource utilization (ISRU) of water by human missions to Mars (Gallagher et al., 2021; Hoffman et al., 2017; I-MIM MDT, 2022; McEwen et al., 2020; Morgan et al., 2021; Phillips-Lander, 2024). The study of VFFs has implications for future mission planning i.e. Mars Life Explorer (MLE); International Mars Ice Mapper Mission (I-MIM) and human exploration on Mars, specifically for assessing the material properties in advance of ISRU and represent a significant window to climate records of Amazonian (~2.9-3.3 Ga) ice and dust cycles (Driver et al., 2024; Michael, 2013).

4.1.2 Study Region

Phlegra Montes (centered on 165°E, 40°N) is a region in northern mid-latitudes that bounds a 1400 km long mountain range northeast of Elysium Mons (**Figure 4.1**). Phlegra Montes spans from 30° N to 50° N, a region in the Martian midlatitudes that expresses the gradation of the presence of subsurface ice from ice-free to ice-rich as latitude increases.

Systems of ridges, troughs, fractures, channels, and dispersed craters mark the northern plains. The largest mountainous region in the northern plains is the north trending, degraded ridges of Phlegra Montes. Phlegra Montes is located northeast of the Elysium rise complex and was formed as a result of tectonic uplift (Early Hesperian – Amazonian) merging with the younger terrain of the Elysium region (Fastook et al., 2011; Greeley and Guest, 1987; Hall et al., 1986). The center of the rise (28°N, 199°W) is near the southern end of Phlegra Montes. There are several west- to west-northwest-trending graben-like lineaments that also cut Phlegra Montes as seen in a valley system infilled by ice at ~40°N (Tanaka et al., 2005).

Moreover, evidence of past glaciation on Mars, as inferred from geomorphic and stratigraphic records (Becerra et al., 2021; Butcher et al., 2023; Hepburn et al., 2020b; Holt et al., 2008), has provided insights into the climatic conditions and mechanisms that governed ice accumulation and deposition, ultimately leading to the formation of glaciated systems or VFFs. The recognition of recent ice ages on Mars (Head et al., 2003; Smith et al., 2016), combined with (1) a more in-depth understanding of climatic driving forces for the last ~20 million years (Laskar et al., 2002), and (2) improved atmospheric general circulation models (GCM) (Fastook et al., 2011, 2008; Levrard et al., 2007; Madeleine et al., 2009) have allowed a more detailed understanding of the recent climate and glaciation history of Mars. In general, periods of higher

obliquity enable the transport of water-ice on the Martian poles to sublimate, transport, and deposit as frost or snow on mid-latitude surfaces (Forget et al., 2006). This results in the accumulation of massive ice and the formation of glaciers and/or ice-rich landforms (Late Amazonian) (Mellon and Jakosky, 1995). By contrast, during periods of low modeled obliquity, ice is removed from the near subsurface equatorward of 60° – 70° latitudes. However, deeper and thicker ice deposits in the mid-latitudes from glaciers may persist longer than shallow or equatorial deposits due to its greater burial depth and insulation, even though low-obliquity generally promotes instability. In Phlegra Montes, the present-day ice is a direct product of climate-driven accumulation, loss, and preservation of ice over multiple obliquity cycles as relicts of a past climatic regime. This change in climate contributes greatly to the geomorphological evolution of VFFs in Phlegra Montes, specifically over the Late Amazonian (~ 300 – 600 Ma to present day).

Varying environmental conditions across this latitude gradient such as topographic relief, altitudinal variation, volcanic activity, and presence of ice-related landforms provide a unique opportunity to investigate this region as a high-value target for cryospheric science and its potential for water-ice as in-situ resource (McEwen et al., 2020). Previous studies in Phlegra Montes (Dickson et al., 2010, 2010; Gallagher et al., 2021; Gallagher and Balme, 2015; Kress et al., 2010; McEwen et al., 2020) have qualitatively substantiated the presence of landform morphologies indicative of subsurface ice. Due to the diverse glacial and periglacial landforms within Phlegra Montes, this region offers a unique window into Mars' climatic history, water-ice distribution and the influence on subsurface ice access and preservation.

4.1.3 Past SHARAD studies of mid-latitude ice

Radar sounding using the Shallow Radar (SHARAD, Seu et al., 2007) offers a unique window to “see” into these ice-rich landforms due to SHARAD’s ability to image subsurface interfaces and provide insights on physical properties such as composition, thickness, ice depth, and spatial distribution of ice within these landforms.

SHARAD is an orbital radar sounder that transmits a 15–25 MHz linearly modulated (“chirped”) pulse. The instrument has a theoretical 15 m vertical resolution in free space, which translates

to $\frac{15}{\sqrt{\epsilon'}}$ m when passing through a medium with a real dielectric permittivity ϵ' , approximately 8 m in water-ice ($\epsilon' \sim 3.15$) (Fujita et al., 2000). After synthetic aperture processing, SHARAD's cross-track resolution is 3–6 km, and the along-track resolution is 0.3–1 km (Seu et al., 2007). SHARAD observations of mid-latitude ice in the 30° to 60° latitude hemispheric belts are mainly concentrated in the northern hemisphere along the topographic dichotomy such as in Deuteronilus Mensae (Baker and Carter, 2023; Baker and Head, 2015; Perry et al., 2023; Petersen et al., 2018c), Utopia Planitia (Campbell and Morgan, 2018; Séjourné et al., 2019; Stuurman et al., 2016), Phlegra Montes (Gallagher et al., 2021; Safaeinili et al., 2009; Steinberg et al., 2025), and Arcadia Planitia (Bramson et al., 2015; Campbell and Morgan, 2018; Ramsdale et al., 2018) as well as in the south around massifs on the rims of Hellas Basin (Grau Galofre et al., 2022; Holt et al., 2008; Karlsson et al., 2015) and Argyre Basin (Berman et al., 2021). These latitudinal bands correlate with the shallow-subsurface stability field for water-ice, where ice could remain metastable just below the surface but would sublime if exposed due to unstable surface conditions (Bramson et al., 2017).

Previous studies of real dielectric permittivity for water-ice reservoirs in the midlatitudes have been proposed to range from $\epsilon' = 2.5$ to 4.5. While not specifically focused on VFFs, but rather subsurface ice, Bramson et al. (2015) initially reported a mean dielectric permittivity of 2.5 ± 0.3 in Arcadia Planitia, while Stuurman et al. (2016) found a mean dielectric permittivity of 2.8 ± 0.8 for a widespread unit in Utopia Planitia, consistent with 50%–85% ice content, with the rest as pore space. The Mars Subsurface Water Ice Mapping (SWIM) project (Morgan et al., 2021) reported a revised dielectric permittivity of 4.0 ± 1.4 for mantling material in Arcadia Planitia, larger than the range of 2.5 ± 0.3 that Bramson et al. (2015) found previously. This indicates a reduced overall ice fraction, suggesting that the material does not consist of massive ice from the surface to the depth of the reflector, but rather a mixture of ice, lithics, and empty pore space. Additionally, in eastern Hellas Basin, Holt et al. (2008) found a dielectric permittivity of 3.2 for several VFFs. Similarly, Baker and Carter, 2023; Morgan et al., 2021; Petersen et al., 2018 calculated dielectric permittivity ranging from 3.2–4.5 for VFFs and mantle material in Deuteronilus Mensae. These studies provide constraints on dielectric permittivity values and ice content estimates for mid-latitude ice deposits, which can serve as analogs for assessing the

composition and structure of VFFs in Phlegra Montes. However, many of these studies focus on isolated sites and do not provide regional-scale constraints on ice thickness or purity.

To address this gap, we present new evidence from the Shallow Radar (SHARAD) sounding experiment on Mars Reconnaissance Orbiter (MRO) (Seu et al., 2007) of glacial systems in the Phlegra Montes region. This is the first SHARAD mapping survey that characterizes the regional ice thickness and subsurface of the entire Phlegra Montes region. The goal of this study is to investigate the distribution of water ice deposits in the Phlegra Montes VFFs and characterize their physical properties using SHARAD. SHARAD subsurface reflections are used to quantify ice thickness and volume for debris-covered glaciers or VFFs and to estimate the thickness of the regolith cover. As all known VFFs on Mars' midlatitudes are debris-covered, we hypothesize that the debris cover in the Phlegra Montes plays a significant role in the region's response to changes in past climate. In this study, we test this hypothesis and provide new regional constraints on dielectric permittivity, debris quantity, ice thickness, and volume that can inform future efforts to evaluate the role of debris cover in VFF stability.

4.2 Methods

4.2.1 Mapping Datasets

This research uses radar data from the Shallow Radar (SHARAD; Seu et al., 2007), imagery from the Context Camera (CTX; 6 m/px; Malin et al., 2007), blended topographic data from the Mars Orbiter Laser Altimeter and High Resolution Stereo Camera (MOLA-HRSC blended DEM, 200 m/px resolution; Smith et al., 2001), and maps from the Thermal Emission Imaging System (THEMIS Night IR; 100m/px; Christensen et al., 2004). CTX and THEMIS were used as basemaps to create shapefiles for nine target glaciers and/or VFFs. Targets were chosen based on their morphology, size, and isolation from other contributing and complex VFFs and topography. CTX and THEMIS provided a holistic and spatial overview of which VFFs we could target in a wide (30°-50°N) latitude range for SHARAD to easily “see”.

For the majority of our analysis and calculations, we use the SHARAD dataset. Primarily, we used the US SHARAD products in the Planetary Data System (PDS). Depth corrected radar images are not available from the PDS, so for each figure with a depth-corrected image, we used data products created from the Co-SHARPS web interface (i.e. QDA). CO-SHARPS access

instructions and the radargrams are available through the US SHARAD Team's website at <https://sharad.psi.edu>. The areoid and ellipsoid datums create artificial differences in topography based on the reference datum that can confuse interpretation, so we used only the PDS products for radar analysis. We compiled and analyzed over 650 intersecting SHARAD radargrams with associated clutter simulations (available on Co-SHARPS or the PDS) of the study site using geophysical interpretation software (SeisWare™) and geographic information system software (i.e., ArcGIS Pro, and JMARS; Christensen et al., 2009).

SHARAD signals reflect the changes and contrasts of the relative permittivity of various materials (i.e. between air-rock or ice-rock). On Mars, real dielectric permittivity values of materials may vary from 1 (atmosphere) to ~12 (dense basaltic lava flows) (Cereti et al., 2007; Heggy et al., 2001). The dielectric value for most non-porous water ice ranges from ~3.0 to 3.2 (Plaut et al., 2007; Brouet et al., 2016; Holt et al., 2008; Plaut et al., 2009; Grima et al., 2009). Values greater than 3.2 generally imply some volume fraction of lithic material (Bramson et al., 2015; Brouet et al., 2016; Carter et al., 2009; Stillman and Grimm, 2011).

We completed clutter analysis to validate the interpretation of subsurface reflectors in our observed power radargrams. “Clutter” refers to reflections from off-target objects or terrain features that may interfere with detection and identification of subsurface reflectors (Choudhary et al., 2016). If a reflection occurs in both the clutter simulation and observed radar image, we assume that it is from the surface and exclude it from further interpretation. Reflections not appearing in simulations are treated as subsurface interfaces and included in our analysis. Due to the complex geometry of the Phlegra Montes topography and the orientation of SHARAD's ground track parallel to the primary mountain chain, radar signals reflect off terrain features that produce abundant clutter.

Another aspect is the identification of radar processing sidelobes. Sidelobes refer to processing artifacts that occur during pulse compression. Sidelobes occur at precise time delays past true reflections and can be identified by their position (at ~0.2 μ s time delay from the first surface return) in time and by their parallel nature with respect to the surface reflection. Together, clutter and side lobes complicate subsurface detection, especially for VFFs, and must be considered

when identifying valid basal reflections. For more information about clutter mitigation, see Smith and Holt (2015).

Another aspect is the identification of radar processing sidelobes. Sidelobes refer to processing artifacts that occur during pulse compression. Sidelobes occur at precise time delays past true reflections and can be identified by their position (at $\sim 0.2 \mu\text{s}$ time delay from the first surface return) in time and by their parallel nature with respect to the surface reflection. Clutter and side lobes can complicate the interpretation of radar data and make it more challenging to detect subsurface reflectors of VFFs, making clutter analysis a fundamental step to confirm a radar basal detection. For more information about clutter mitigation, see Smith and Holt (2015).

4.2.2 Depth Correction

The real portion of the dielectric permittivity of ice is ~ 3.15 , which slows light propagation and delays subsurface reflections (Gough, 1972; Holt et al., 2008; Plaut et al., 2009; Grima et al., 2009). To account for this, a "depth correction" algorithm is applied via the CO-SHARPS Processing Boutique (Putzig et al., 2016), which adjusts the radar wave velocity below the ground surface, using the MOLA gridded data, to that of ice using the formula (**Equation 1**):

$$\varepsilon' = \left(\frac{ct}{2d} \right)^2 \quad (1)$$

or

$$d\sqrt{\varepsilon'} = \frac{ct}{2} \quad (2)$$

where ε' is the bulk dielectric permittivity of the material, c is the speed of light ($300 \text{ m}/\mu\text{s}$ in a vacuum), t is the two-way travel time of the radar signal (time between the surface return and the subsurface reflection, usually in microseconds), and d is the thickness of a unit or depth to the subsurface feature as measured from the MOLA surface topography.

To test the permittivity of each feature, we use depth correction at various values of ε' (3.0, 3.1, 3.15, 3.2, 3.3, 3.4, 3.5, 3.6, 4.0) to constrain the thickness and purity of the ice. We do this multiple ways by ensuring that the geometries of the detected reflections align with the surrounding terrain that extends beneath the VFFs (i.e. using MOLA in **Appendix; Table 1**), as subsurface interfaces for some LDAs follow pre-existing topography (e.g. Holt et al., 2008; Plaut

et al., 2009). For one portion of the estimation, we assume a flat subsurface geometry as it simplifies radar signal interpretation, minimizes depth calculation uncertainties, and provides a first-order approximation in the absence of a detailed basal topography constraints. For another we project the regional slope in towards the LDA base. This quantitative process is further shown in **Appendix, Table 1**. After we apply this correction, the radar data better represent the geometry of subsurface reflectors relative to the surface topography (i.e. physical relationships between surface and subsurface features), allowing for better interpretation of basal features. (Gallagher et al., 2021) found a value of $\epsilon' = 3.6$; which we test.

While the true physical system consists of three layers (i.e. supraglacial lag, ice, and substrate), we adopt a two-layer model of lag deposit on top of the ice for the purpose of estimating the bulk dielectric permittivity, assuming the basal interface (ice-substrate) acts as a reflecting boundary. In our first method, we follow the methods of Gallagher et al. (2021) to determine the best fitting ϵ' for the bulk material.

$$\epsilon_b' = \left(\frac{ct_t}{2d_b} \right)^2 \quad (3)$$

Where the bulk permittivity (ϵ_b') is calculated from the total time (t_t) between surface and basal reflections and the bulk total thickness (d_t). The bulk permittivity can be broken up into individual components in **Equation 4**

$$\epsilon_b' = \left(\frac{c(t_1 + t_2)}{2(d_1 + d_2)} \right)^2 \quad (4)$$

Where t_1 is the two-way travel time for the radar signal through the top layer (debris), t_2 is the two-way travel time for the radar signal through the bottom layer (ice) to the substrate, d_1 is the thickness of the top layer (rock), and d_2 is the thickness of the bottom layer (ice) (Figure 2), so that d_t from equation 2 is thus the sum of d_1 and d_2 , and t_t is the sum of t_1 and t_2 . All of t_1 , t_2 , d_1 , and d_2 are unknown; however, we know that $t_2 = t_t - t_1$ and that $d_2 = d_t - d_1$. **Equation 4** can be rewritten as

$$\epsilon_b' = \left(\frac{\frac{ct_1}{2} + \frac{ct_2}{2}}{d_1 + d_2} \right)^2 \quad (5)$$

And substituting **Equation 2** into **Equation 5** yields

$$\varepsilon'_b = \left(\frac{d_1 \sqrt{\varepsilon'_1} + d_2 \sqrt{\varepsilon'_2}}{d_1 + d_2} \right)^2 \quad (6)$$

or the equation for the bulk dielectric permittivity (ε'_b) of a two-layer system in which each component is broken out. ε'_1 is the dielectric permittivity of the top layer (supraglacial deposit; for this we assume to be lithic plains forming material from Mouginot et al., 2012 or unconsolidated regolith and use $\varepsilon'_1 = 8$ or 5.5), and ε'_2 is the dielectric permittivity of the bottom layer (we assume to be nearly pure ice and use $\varepsilon'_2 = 3.15$, based on previous studies (Evans, 1965; Grima et al., 2009). We recognize that variations in composition (e.g., grain size, porosity, and dust content) can introduce uncertainty from our assumptions. We have included a calculation in Table 1 (Appendix) to show that if we assume supraglacial debris is made of basaltic clasts (generally $\varepsilon' = 9$) with 30% bulk porosity, the bulk permittivity then becomes $\varepsilon' \approx 5.5$. Important note in advance of the results: even using $\varepsilon'_1 = 5.5$ instead of 8.0, there is not a significant change in the results for lag thickness or total ice volume. The change in ice volume is smaller than the last significant digit of ice volume in the individual glacier and in the full regional volume.

To date, no SHARAD studies have identified a reflection from supraglacial debris, and the supraglacial debris are interpreted to be too thin to create an independent reflection at SHARAD's resolution (Baker and Carter, 2019; Plaut et al., 2009). Therefore, as a first constraint, we can use the non-detection of radar reflections below the surface to learn that the debris are not thick enough to create these reflections. Past studies have assumed SHARAD's vertical resolution in free space to be 15 m, which is only valid if no windowing is applied during pulse compression. The Hann windowing function used in most SHARAD products (e.g., FPA, FPB, QDA) effectively degrades the vertical resolution to ~ 30 m in free space and ~ 20 m in water ice (Nunes and Phillips, 2006; Heinzel et al., 2002). However, we still use 15 m as a baseline reference for consistency with previous studies that assumed a bulk dielectric permittivity closer to $\varepsilon' = 4$. While we acknowledge that 15 m may be an underestimate on the uncertainty, this is not a true constraint because all of the calculated debris thicknesses we report in the Section 3 are < 10 m. Further, we include it as a comparative baseline and explicitly discuss the implications and uncertainties associated with this assumption in Section 2.4. Once

d_1 is found, we can readily calculate d_2 from our measurements of d_t . The thickness (d_t) of each VFF refers to the bulk thickness including both the lag and ice components. When we estimate the volume of ice for each glacier, we use only the ice thickness (d_2) calculated d_t minus lag thickness (d_1).

We assume a simplified model where the radar signal interacts with the layers as if they are stacked in series (**Figure 4.2**) as also used by Berman et al. (2021) and Gallagher et al. (2021).

$$d_1 = d_t \left(\frac{\sqrt{\epsilon_b'} - \sqrt{\epsilon_2'}}{\sqrt{\epsilon_1'} + \sqrt{\epsilon_2'}} \right) \quad (7)$$

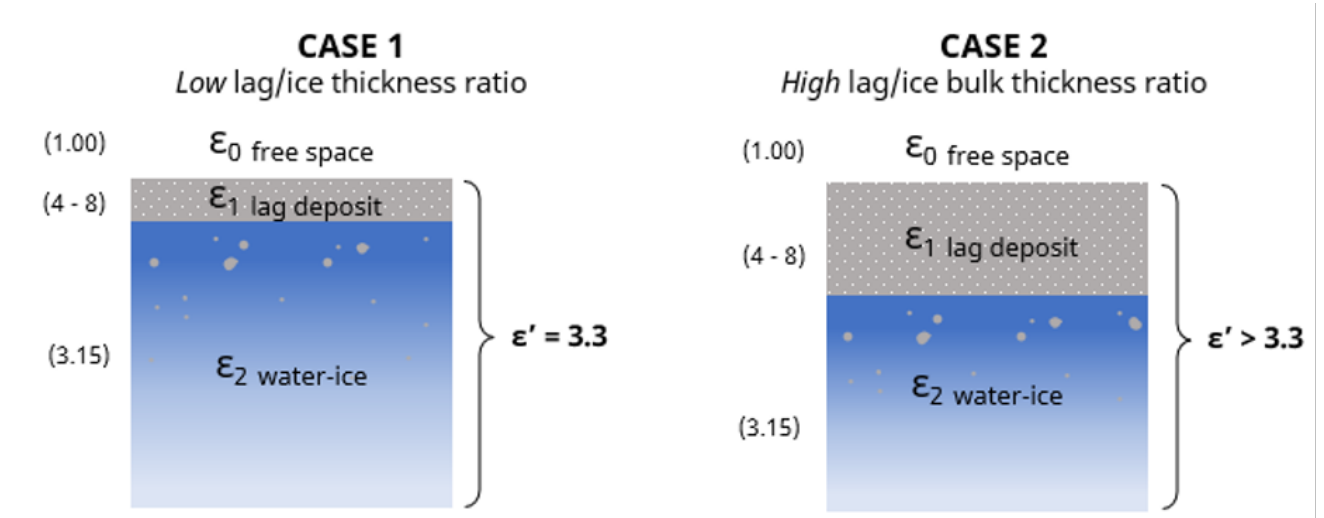


Figure 4.2. Model schematic of the effect of the thickness of lag to the bulk dielectric permittivity for a two-layer system. Case 1 exhibits a low lag to ice ratio and therefore a bulk dielectric value of 3.3, while Case 2 (this study) proposes a higher lag to ice ratio for a bulk dielectric of equal to or more than 3.3 signifying a thicker lag deposit.

In order to calculate the maximum thickness of d_1 and d_2 and bulk dielectric value, we used the (surface and subsurface reflector) two-way time travel from SHARAD and extrapolated the surface elevation from MOLA. *MOLA high* represents the surface of the glacier and *MOLA low* represents the surface of the flat plains adjacent to the glacier (extrapolated version of the glacier base). These values are then able to complete Equation 7 to provide a more accurate dielectric estimation (all values reported in Table 1 in Appendix).

Least square analysis to determine bulk dielectric permittivity of Site 1: Main LDA

To better constrain the bulk dielectric permittivity of Site 1’s main LDA, we performed a least-squares analysis, a statistical method that minimizes the sum of the squared differences between observed and predicted values. In this case, we compared the radar-derived surface picks with two idealized basal geometries: (1) a flat horizontal base extrapolated from the surrounding terrain, and (2) a sloped base that continues the local surface gradient beneath the glacier. We then determined which permittivity value best corrected the radar time delay to match each basal assumption.

To calculate volume, we use the mean of the depth-converted picks multiplied with their mapped area from the CTX-mapped boundary shapefile. In our results, for every site we report the average thickness of the glacier, the maximum ice thickness ($d_{2\max}$) and maximum lag ($d_{1\max}$) across the different SHARAD observations, as well as the glacier water-ice volume excluding the volume of the supraglacial debris. While average thickness is influenced by ground track location and edge effects, we include it for comparative purposes across sites. Bulk ice content, also referred to as ice purity, was calculated as the ratio of ice thickness to total subsurface thickness [$\text{ice}/(\text{ice} + \text{lag})$], using values derived from the dielectric permittivity and lag assumptions reported in Section 3 (also in Table 1 Appendix).

4.2.3 Loss tangent calculation

The loss tangent is the ratio of imaginary and real parts of the complex dielectric permittivity of the material, which physically represents the attenuation, or non-scattering power loss, of the radar wave as it traverses through a medium. Previous studies of LDAs across the midlatitude regions of Mars present loss tangent values from 0.001 and 0.005 (Baker and Carter, 2023; Campbell and Morgan, 2018; Petersen et al., 2018c; Plaut et al., 2007; Watters et al., 2007).

For loss tangent calculations, we retrieved MOLA topography from the SHARAD-associated files on the PDS and imported them into Seisware™. We then “snapped” the MOLA topography to bright reflectors (i.e. nadir surface return) and extracted the surface signal power. This provided the surface reflection strength. We then measured the along-track power of subsurface reflectors in the Phlegra Montes region. The power is then used to characterize the signal sensitivity in terms of the power loss (dB) and time delay (μs) similar to methods used in recent studies (Bharti et al., 2022; Campbell and Morgan, 2018). Following their methods, we calculate

the loss tangent, which corresponds to the attenuation between the surface and subsurface reflector using **Equation 8**:

$$\tan(\delta) = \sqrt{\left[\left(\frac{\lambda(\ln L)}{4\pi c \Delta t} \right)^2 + 1 \right] - 1} \quad (8)$$

where $\tan(\delta)$ is the loss tangent, λ is the SHARAD wavelength in free space (ranging from 12 to 20 m, but assumed a constant 15 m for our work), and $\ln L$ is the natural logarithm of the power loss (L) per unit of time (Δt). Because loss tangent is a function of power loss and time, multiple points at various time delays are required to find a slope that represents the loss tangent. To find that slope, we create linear regression plots and determine their slope.

4.2.4 Uncertainty considerations

Our calculations include several sources of uncertainty. First, the two-way travel time uncertainty of the surface and subsurface reflectors is accounted for using a combination of SHARAD's instrument resolution and the standard deviation of ice thickness measurements. The minimum vertical time delay uncertainty for SHARAD is 37.5 ns (1 sample), corresponding to ~3 m in water ice. Since this uncertainty applies to both the top and basal reflectors, we adopt a total two-way time uncertainty of 112.5 ns (~9.5 m in water ice) plus the standard deviation of thickness measurements i.e. *thickness* $\pm$ (9.5 + *std dev*) as reported in **Sections 3.3.1 to 3.3.9**.

Second, uncertainty due to the subjectivity of reflector picking is considered. The identification of reflectors inherently involves some interpretation on Seisware due to its peak-following algorithm; however, previous assessments and independent verification suggest that such subjectivity does not introduce meaningful changes to the results. To ensure consistency, the reflector picks were verified by both authors and, when necessary, repicked to minimize interpretation biases. As part of our methodology, we report the instrument-derived uncertainties associated with SHARAD data to ensure transparency.

Third, uncertainty related to the topographic irregularity of the surrounding terrain is acknowledged. While local variations in topography can introduce error, we do not expect this to significantly affect our results, as the applied radar uncertainty already accounts for these variations. We note that our approach relies on a single MOLA point (extracted from the MOLA

interpolated product) for elevation (**Table 1 in Appendix**), which introduces potential errors not explicitly quantified in our calculations. However, even assuming an error of ± 10 m in surface elevation, near the limit of SHARAD's vertical resolution in ice, the resulting impact on volume estimates is minimal relative to the total volumes involved (on the order of km^3). For example, a 10 m uniform offset over a 1 km^2 area would correspond to a volume error of only 10^7 m^3 , a minor amount when compared to the glacier volume of $\sim 10^{11} \text{ m}^3$. The impact of uncertainty is even smaller when assuming that the deviation is not a uniform offset but more randomly distributed around 0 m.

Moreover, while MOLA cannot detect the glacier base, we estimate a basal surface by assuming it has a uniform slope (e.g. we consider that it could be sloped following the surrounding terrain beyond the glacier toe or perfectly horizontal: in Table 2 Appendix). These assumptions are necessary due to the lack of direct basal observations and do introduce uncertainty as different basal geometries can yield varying depth estimates and dielectric permittivity values. For example, assuming a flat base versus a sloped base can shift ϵ' by up to 0.25, which directly affects derived lag thickness and ice volume. We report the ranges of volumes calculated based on the two basal assumptions.

Lastly, uncertainty in the loss tangent is estimated using the standard error of the linear regression between power loss and two-way travel time (TWT). This uncertainty reflects the combined effects of surface and basal power measurement variability, TWT picking uncertainty, and unquantified spatial variability in roughness and composition along the radar profile. The range defined by the standard error is used to determine upper and lower bounds on the loss tangent, and is reported alongside the results in **Section 4.3.4**.

4.3 Results

4.3.1 Regional ice and regolith thickness

We found several radargrams across Phlegra Montes with strong subsurface reflections that are not adversely affected by topographically induced radar clutter and rate the reflections as “high confidence”. After converting to thickness (next subsection) The overall result of this survey is an apparent radar ice thickness map of the several sites in the region of Phlegra Montes (**Figure 4.3**). We use this map to estimate regional ice distribution and total ice volume.

Using the two-layer model outlined in the *Methods*, we calculate a maximum thickness (d_{lmax}) of the lag or supraglacial debris for each site ranging from 1-8 ma. The average bulk thickness (d_{tave}) of all mapped VFFs with positive detections in eight out of nine study sites ranged from 135-220 m, spanning a wide latitudinal range (30-50° N) with the maximum apparent ice thickness reaching up to 400 ± 14 m. One basal non-detection site (Site 3) was not included in this calculation.

4.3.2 SHARAD data

Numerous SHARAD tracks in our survey cross specific targeted regions (i.e. isolated LDAs, large VFF complexes) and exhibit strong subsurface reflections. Sites in the following subsections were grouped and are presented in the next sections from lower to higher latitude, with the site in the lowest latitude presented first. In total, there are nine notable regions (**Figure 4.3**): *Site 1: Main LDA*, *Site 2: Main Composite System*, *Site 3: Side LDA E-Flank (Null Detection)*, *Site 4: Double LDA NE-Flank (Null Detection)*, *Site 5: LDA-Composite System*, *Site 6: N-LDA 1*, *Site 7: N-LDA 2*, *Site 8: N-LDA 3*, and *Site 9: N-LDA 4*.

4.3.2.1 Main LDA (Site 1)

Numerous SHARAD tracks in our survey cross specific targeted regions (i.e. isolated LDAs, large VFF complexes) and exhibit strong subsurface reflections. Sites in the following subsections were grouped and are presented in the next sections from lower to higher latitude, with the site in the lowest latitude presented first. In total, there are nine notable regions (**Figure 4.3**): *Site 1: Main LDA*, *Site 2: Main Composite*

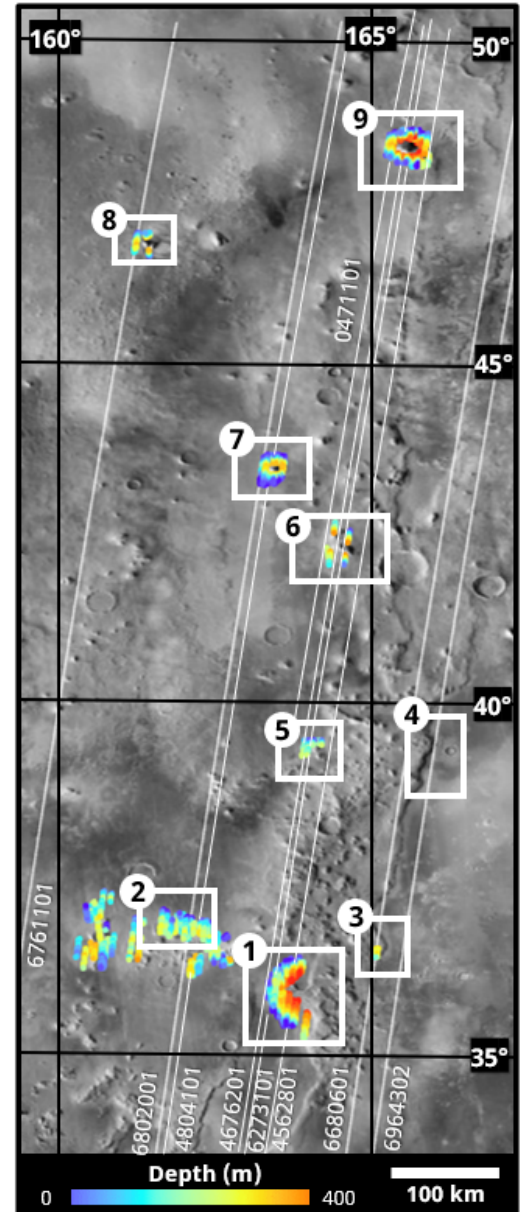


Figure 4.3. An apparent depth map of selected study sites as a proxy for ice thickness in Phlegra Montes. The main study sites are numbered in white boxes highlighted in the Results. White lines show the individual SHARAD tracks.

System, Site 3: Side LDA E-Flank (Null Detection), Site 4: Double LDA NE-Flank (Null Detection), Site 5: LDA-Composite System, Site 6: N-LDA 1, Site 7: N-LDA 2, Site 8: N-LDA 3, and Site 9: N-LDA 4.

The largest and southernmost LDA in Phlegra Montes is at 35.7°N, 163.9°E and was studied by Gallagher et al. (2021) and Steinberg et al. (2025). Due to its area and lobate shape, this feature is identifiable both in visible imagery and radargram cross-sections. To estimate the bulk dielectric value, we tested $\epsilon' = 3.15, 3.2, 3.3, 3.4, 3.5$, and 3.6 . The maximum bulk thickness of this LDA is 400 m compared to the maximum potential thickness of 500 m found by Gallagher et al. (2021). The average bulk thickness across this LDA is 216 ± 12 m (2 m of standard deviation + 9.5 m of SHARAD resolution two-way time uncertainty; equivalent to 7.8% uncertainty in the thickness) with a maximum lag thickness ($d_{l \max}$) of 7 m and a maximum ice thickness ($d_{2\max}$) of 392 m across the different observations (**Table A1 in the Appendix**). We do not report uncertainties for the maximum values as they represent single-point estimates rather than statistical averages, and assigning uncertainty could imply a level of precision that is not supported. The total water-ice volume approximation of $\sim 3.7 \times 10^{11} \pm 2.9 \times 10^{10} \text{ m}^3$ excluding the volume of the supraglacial debris. An example radargram of this detection is shown on **Figure 4.4**.

In contrast to the $\epsilon'=3.6$ that Gallagher et al., (2021) found and Bain et al. (2022) used, we reinterpret the data and find that $\epsilon'=3.3$ is more consistent with the data (**Figure 4.4**) because $\epsilon' = 3.2$ fits better with the known constraints of a thin overburden that does not provide a detectable subsurface interface above the ice (**Equation 7**). This value is used uniformly in the radargram figures for depth correction and comparison purposes. However, site-specific bulk dielectric values, derived from MOLA elevations and TWT, are provided in **Table A1 in the Appendix**, for quantitative analysis.

Due to the high coverage of SHARAD tracks crossing this large area, confirmed basal reflections for Site 1 are mapped with a high confidence level. With this, we generated and interpolated an apparent depth map using ArcGIS Pro's Inverse Distance Weighting (IDW) to visually illustrate the ice thickness of this LDA using $\epsilon'=3.3 \pm 0.25$ and subtracting the lag thickness.

We follow the same methods for each following site. Because the following sites (sections 4.3.3.2 – 3.3.9) do not offer the same, flat geometry (most are on the flanks of massifs or are perched), we assume that all VFFs in Phlegra Montes have the same bulk dielectric value ($\epsilon_b' = 3.3 \pm 0.25$). This is consistent with the global results from Steinberg et al. (2025) who found a narrow range of LDA ϵ' values across five widely dispersed sites across the globe between 3.05 and 3.4.

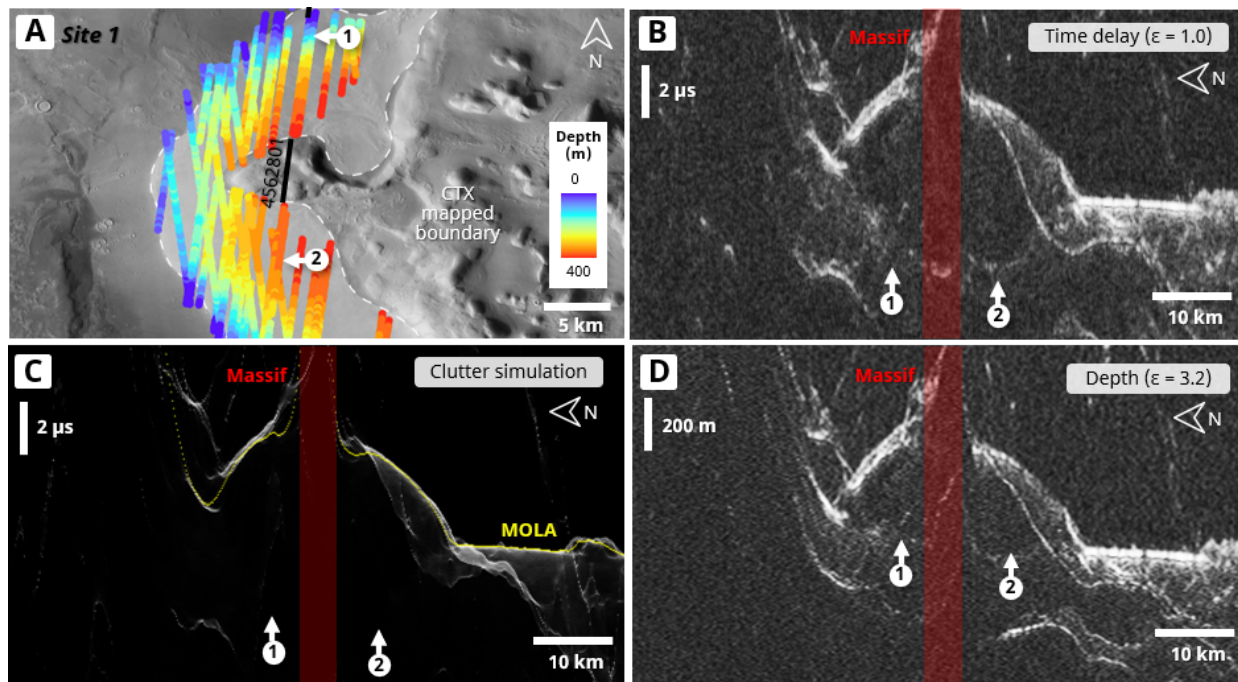


Figure 4.4. Site 1 (A) SHARAD track 4562801 highlighted in white with the locations of subsurface reflectors. (B) Power radargram of the largest LDA in Phlegra Montes (Gallagher et al., 2021) with subsurface reflectors outlined as 1 and 2; (C) Clutter simulation of (B) showing the absence of clutter near the possible subsurface reflections; (D) Depth corrected power radargram ($\epsilon = 3.2$) following the methods of Gallagher et al. (2021). When we use the same dataset as them, in which depth corrections tools are available and referenced to the ellipsoid, we find that 3.2 is not the best bulk dielectric value. However, when we calculate the ϵ' using the aeroid (for which we do not have access to a depth correction image generator) we find that $\epsilon' = 3.6$ is a better value. Depth correction provides more accurate subsurface feature depths by compensating for signal travel time variations. Rather than a “bowl”-shaped subsurface morphology, an improved geometry is seen in (D) as a nearly flat topographic base.

4.3.2.2 Main Composite System (Site 2)

We report the first detection of a composite glacial system (36.58°N, 162.43°E) with undulating, terraced topography to be detected by SHARAD. This detection consists of two subsurface terraces going from higher to lower topography that are hypothesized to be ice-rich due to the presence of LDAs, LVFs, and other VFFs in this composite system. Because the geometry of the basal reflection of the terrace does not correspond to a horizontal plane, as in Site 1, we had to

make an assumption about the bulk dielectric and choose a value of $\epsilon' = 3.2$ to match Site 1. This choice was based on the need for consistency in comparison, though variations due to differences in morphology and composition may exist. We then follow on our measurements and calculations using **Equation 7**. With the depth correction at $\epsilon'=3.2$ using the ellipsoid-derived data product, we are still able to see a good correction that matches the surrounding terrain (**Table A1**). This composite system geomorphologically consists of lineated valley fills, concentric crater fill, and LDAs merging around massifs forming this ice field. The average thickness of this LDA at this northeastern portion, which covers approximately a quarter of the composite system's area is 159 ± 13 m (8.1% uncertainty) with a maximum lag thickness ($d_{1 \text{ max}}$) of 4 m and a maximum ice thickness ($d_{2 \text{ max}}$) of 199 m. The total water-ice volume approximation of $\sim 2.3 \times 10^{11} \text{ m}^3 \pm 1.9 \times 10^{10} \text{ m}^3$ excluding the volume of lag. An example radargram of this detection is shown on **Figure 4.5**. Typically, terrestrial debris-covered glaciers display similar basal reflections, following undulating subsurface topography on terraced bedrock (Anderson et al., 2018; Dickson et al., 2008; Dutrieux et al., 2014; Kane et al., 1973; Mahaney et al., 2007; Robl et al., 2015). To characterize this composite system further, a depth map (**Figure 4.5A**) was generated as a proxy to apparent ice thickness using detected subsurface reflectors from these tracks (**Appendix; Table A1**).

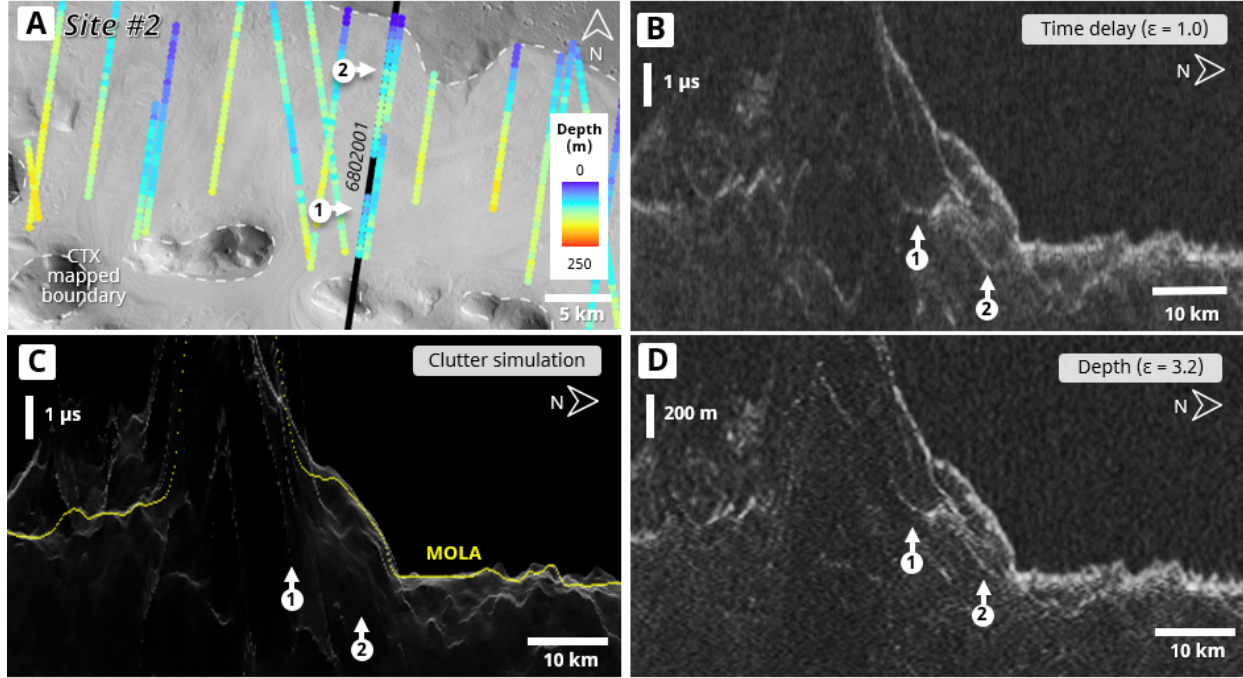


Figure 4.5. Site 2 (A) SHARAD track 6802001 highlighted in white with the locations of subsurface reflectors; (B) Power radargram of the first evidence of terraced subsurface topography on a composite system on Mars with subsurface reflectors outlined as 1 and 2; (C) Clutter simulation of (B) showing the absence of clutter for two terraces; (D) Depth corrected power radargram ($\epsilon = 3.2$), adjusting for signal travel time to present more representative subsurface depth, show the perched and composite nature of Reflectors #1 and #2.

4.3.2.3 Mixed Detections of Side LDA on Eastern Flank (Site 3)

Along the eastern side of the Phlegra Montes mountain range (36.68°N , 165.17°E) there are two LDAs (**Figure 4.6**). Five SHARAD observation cross these LDAs, but only the southern LDA (Reflector #2 in **Figure 4.6**) has positive subsurface reflections. For the northern LDA, most reflections coincide with clutter as shown by the example of Reflector #1 in **Figure 4.6**, and we are unable to confidently assess the thickness or physical properties. We categorize the northern LDA as a null detection.

Interestingly, we observe that Reflector #2 in the southern LDA is topographically higher than Reflector #1. We consider it to be a perched glacier, topographically above the planes to the south and east, terminating just before the southern valley wall. Given this geometry, we interpret this LDA to be similar to terrestrial hanging valleys formed by tributary glaciers. We identify a positive SHARAD reflection for the southern LDA using $\epsilon' = 3.2$ (**Table A1 in Appendix**) with an average thickness of 227 ± 16 m (7.0% uncertainty), a maximum lag thickness ($d_{l \text{ max}}$) of 5 m and a maximum ice thickness ($d_{i \text{ max}}$) of 280 m. The water-ice volume

estimate for the southern LDA is calculated to be $\sim 2.35 \times 10^{10} \text{ m}^3 \pm 1.6 \times 10^9 \text{ m}^3$ excluding the volume of lag. This is the first detection of a perched glacier using SHARAD on Mars. (**Figure 4.6; Table 1A**).

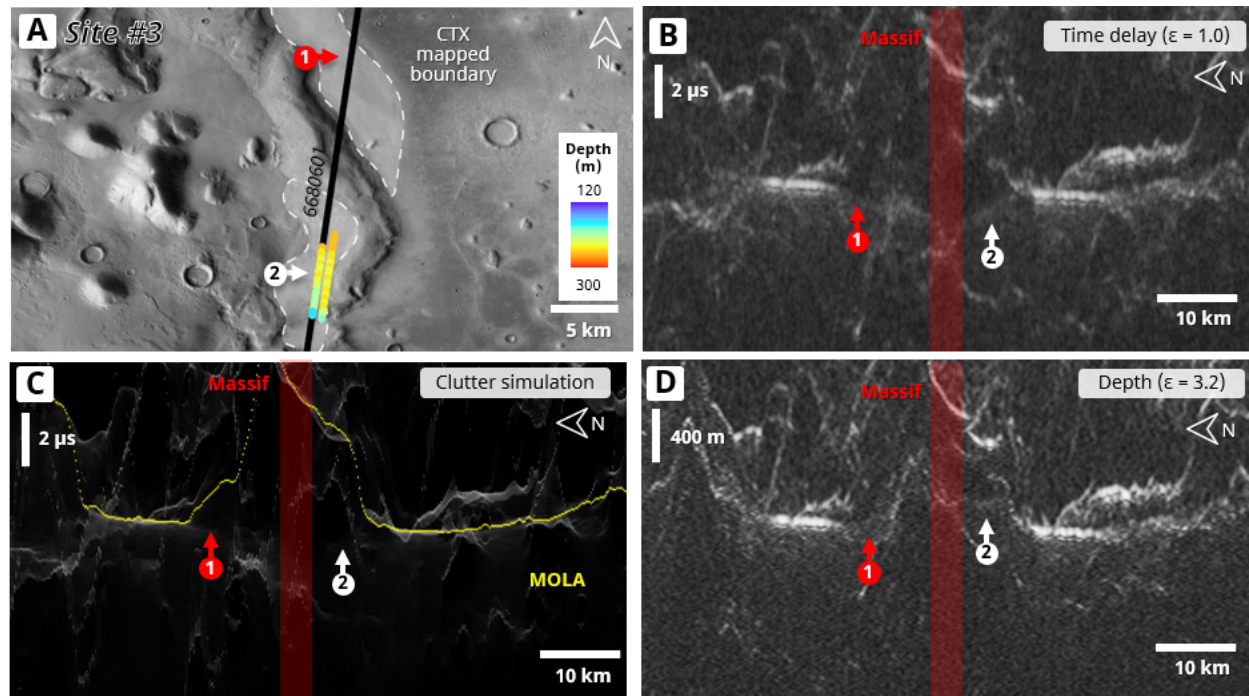


Figure 4.6. Site 3 (A) SHARAD track 6680601 highlighted in black with the locations of possible detections that coincide with clutter. (B) Power radargram of two LDAs on the eastern flank of the valley labeled as Null Reflector # 1 (in red) and Reflector #2; (C) Clutter simulation of (B) showing the absence of clutter near Reflector #2 (in white); (D) Depth corrected power radargram ($\epsilon = 3.2$), showing the elevated base of this ice mass as Reflector #2.

4.3.2.4 Null Detections of Double LDA Northeastern Flank (Site 4)

Another pair of double LDAs exist north of Site 3 on the eastern flank of Phlegra Montes (38.96°N , 165.92°E), and three SHARAD observations cross these LDAs (Table 1).

Unfortunately, these two LDAs flank the eastern slope of the mountain range that trends NE-SW, parallel to the orbit of MRO and SHARAD, which enhances across track clutter (**Figure 4.7**). Thus, similar to Site 3, SHARAD data here are complicated by clutter at time delays where we would expect reflections, and we categorize these LDAs as a null detection.

To attempt to mitigate the clutter, we compared the left and right clutter simulations to further analyze geometric effects of the adjacent mountain range and confirmed that due to the massif wall, there is too much clutter at the expected time delay of a basal reflection to confirm reflections of the two LDAs. We do not estimate the bulk properties of these glaciers.

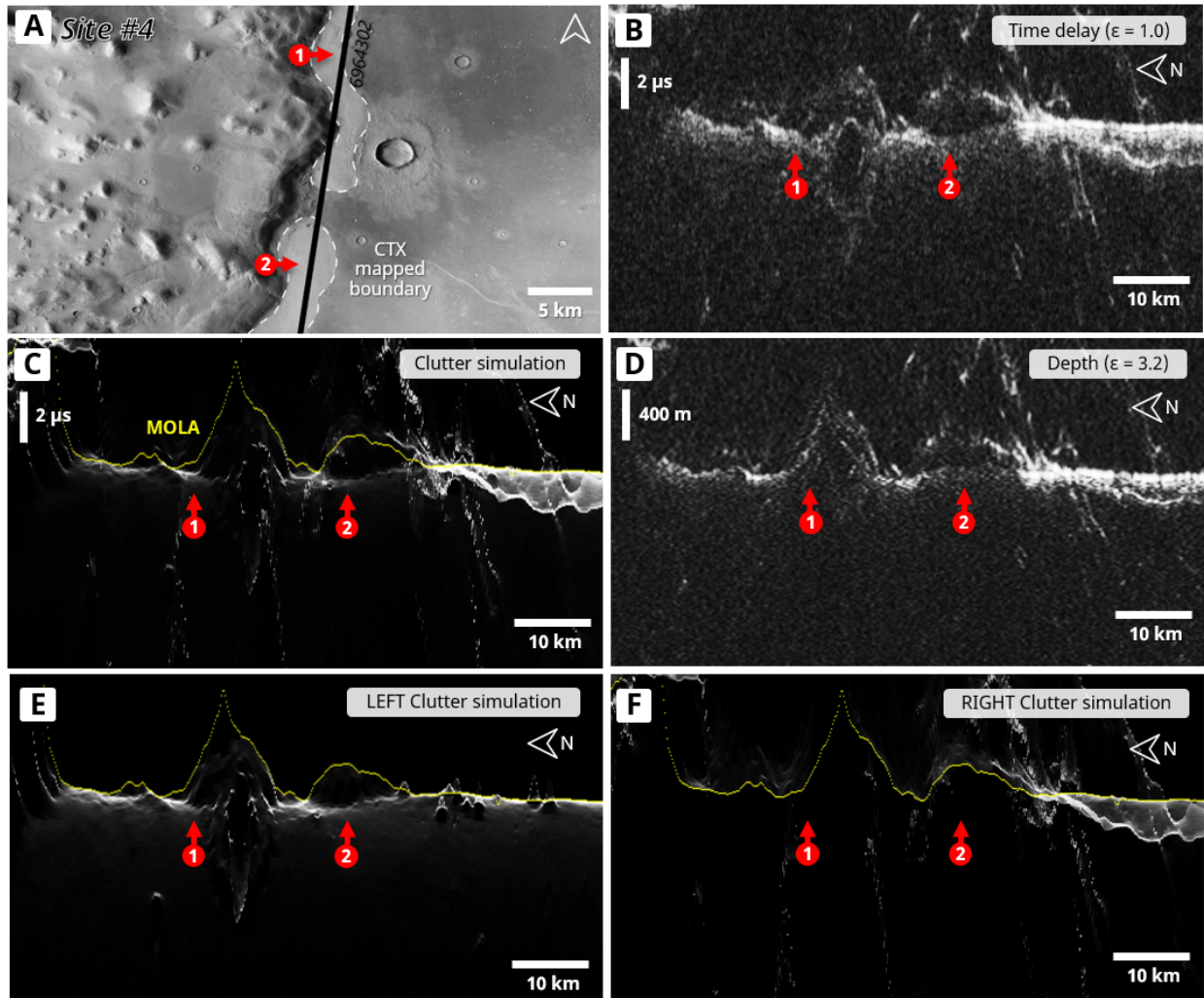


Figure 4.7. Site 4 (A) SHARAD ground track 6964302 annotated with red arrows to show the locations of regions that correspond to possible detections in the subsurface. (B) Power radargram of two candidate LDAs outlined as 1 and 2; (C) Clutter simulation of (A) showing the presence of clutter exactly where a subsurface reflection is expected; (D) Depth corrected power radargram ($\epsilon' = 3.2$), showing how these reflections could indicate a glacier base that is continuous and co-planar to the nearby terrain. (E) and (F) show SHARAD clutter

4.3.2.5 LDA in a second Composite System (Site 5)

This detection is situated within another composite system, with multiple, connected LDAs and LVFs, in Phlegra Montes (39.20°N, 164.°E). This positive detection is positioned towards the northmost edge of the composite system (**Figure 4.8; Table A1**) which made it a more suitable location for SHARAD as it is more isolated from complex topography. The average bulk thickness of this LDA in this second Composite System is 142 ± 14 m (9.9% uncertainty) with a calculated maximum lag thickness ($d_{1\max}$) of 5 m and a maximum ice thickness of ($d_{2\max}$) 199 m.

The total water-ice volume was also calculated to be $\sim 4.9 \times 10^{10} \text{ m}^3 \pm 4.8 \times 10^9 \text{ m}^3$ excluding the volume of supraglacial debris.

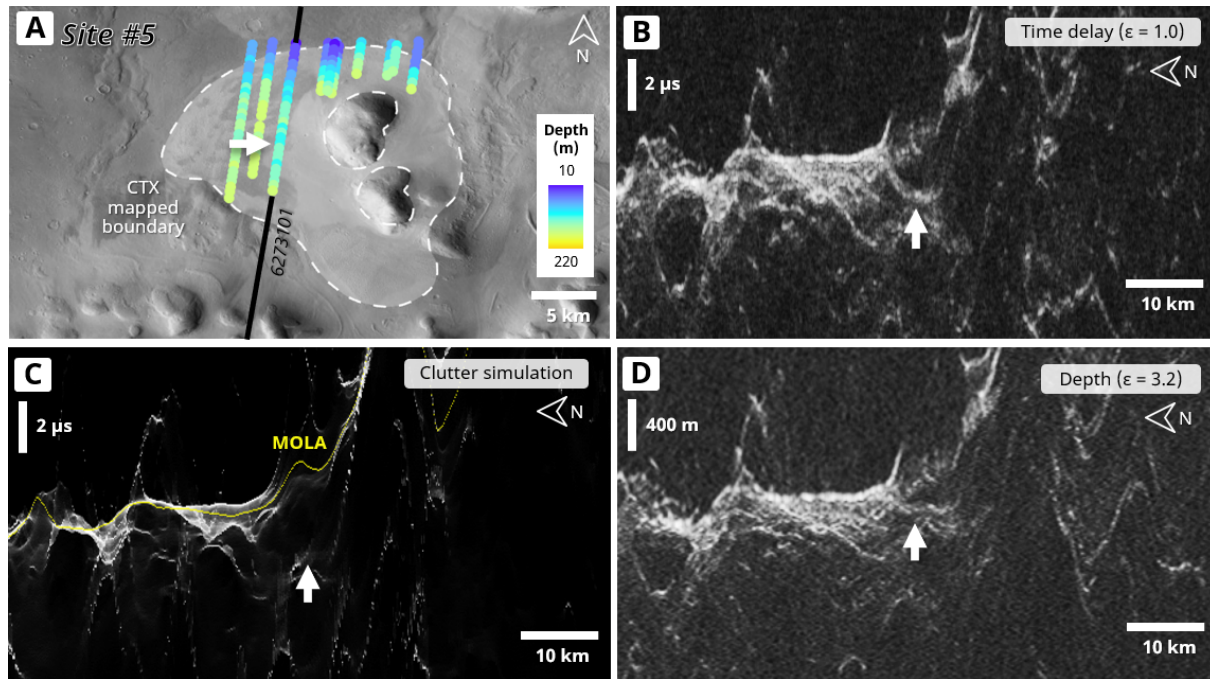


Figure 4.8. Site 5(A) SHARAD ground track 6273101 highlighted in black with colours indicating the location of subsurface reflectors. (B) Power radargram at Site 5 with the subsurface reflector indicated by a white arrow; (C) Clutter simulation of (B) showing the absence of clutter at the same time delay; (D) Depth corrected power radargram ($\epsilon = 3.2$). As in the other cases, $\epsilon' = 3.2$ for the ellipsoid data product does not quite bring the subsurface reflector parallel with the surrounding terrain.

4.3.2.6 North LDA #1 (Site 6)

This subsurface detection is of a Composite System draping two high relief massifs with one reflector detected in the north and one in the south (40.97°N , 164.37°E ; **Figure 4.9**). This is the first LDA in our study north of 40° . The average thickness of this LDA in this Composite System is $203 \pm 16 \text{ m}$ (7.9% uncertainty) with a maximum lag thickness ($d_{1\text{max}}$) of 7 m and a maximum ice thickness ($d_{2\text{max}}$) of 320 m. The total water-ice volume is $\sim 1.8 \times 10^{11} \text{ m}^3 \pm 1.4 \times 10^{10} \text{ m}^3$, excluding the volume of lag. Morphologically, we observe that the massif heads are less exposed than those further south and that the apron of the glacier is gradually blending in with the surrounding “smooth” mantle terrain, making it more difficult to determine VFF boundaries as we move north.

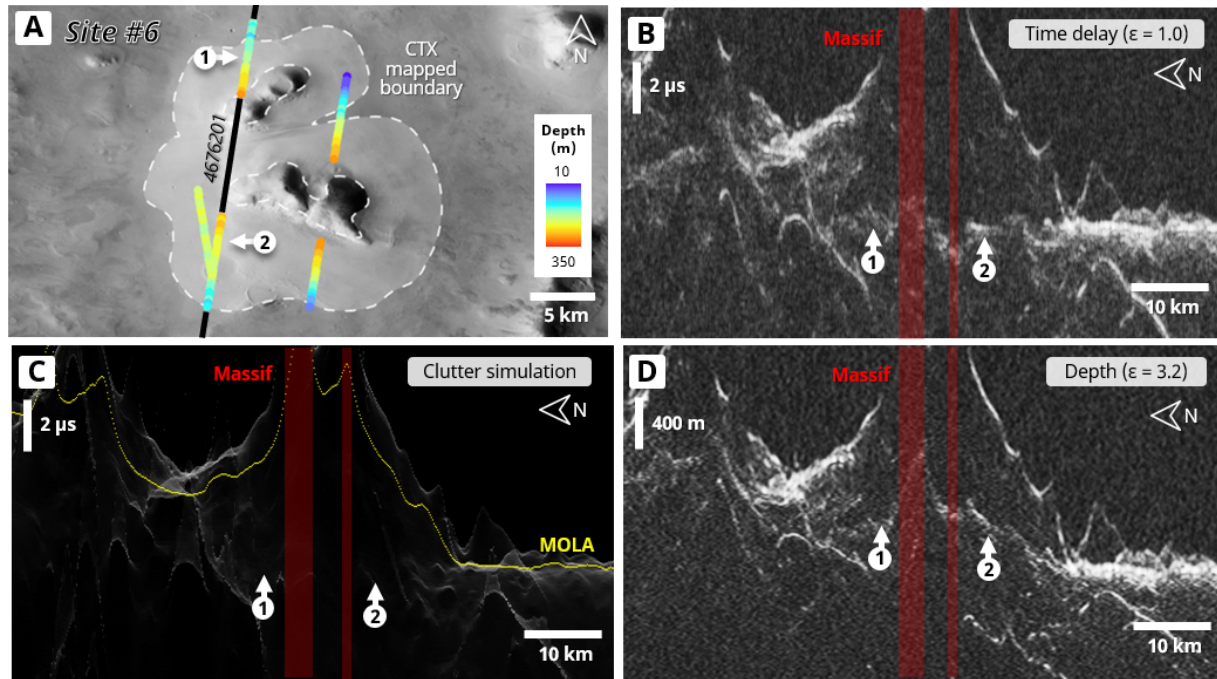


Figure 4.9. Site 6 (A) SHARAD ground track 4676201 highlighted in black with colours indicating the locations and depths of subsurface reflectors over the fourth largest LDA in Phlegra Montes. (B) Power radargram corresponding to the black line in A with subsurface reflectors indicated as 1 and 2; (C) Clutter simulation of (B) showing the absence of clutter at the time delay of the subsurface reflection; (D) Depth corrected power radargram ($\epsilon = 3.2$).

4.3.2.7 North LDA #2 (Site 7)

This LDA is located at 43.2°N , 163.59°E and it is defined by an apron draping a single mound. This morphology is typical for most LDAs in the mid-latitudes (Levy et al., 2014; Plaut et al., 2009) and is the easiest to detect amongst all LDA morphologies. Reflectors are detected in the northern and southern portions of the LDA. We are careful here to only map the LDA thickness where ice is present and not over the mound fully, as subsurface reflectors underneath the exposed mountain must be clutter – in other words, the glacier landsystem is not continuous underneath the surface mound bedrock (**Figure 4.10**). The depth corrected radargram shows this with the middle of the mound in the subsurface being corrected to match the surface topography. The average thickness is 136 ± 14 m (10% uncertainty) with a maximum lag thickness ($d_{1\text{max}}$) of 5 m and a maximum ice thickness ($d_{2\text{max}}$) of 280 m. The total water-ice volume was calculated to be $\sim 6.7 \times 10^{10} \text{ m}^3 \pm 6.9 \times 10^9 \text{ m}^3$ excluding the lag volume.

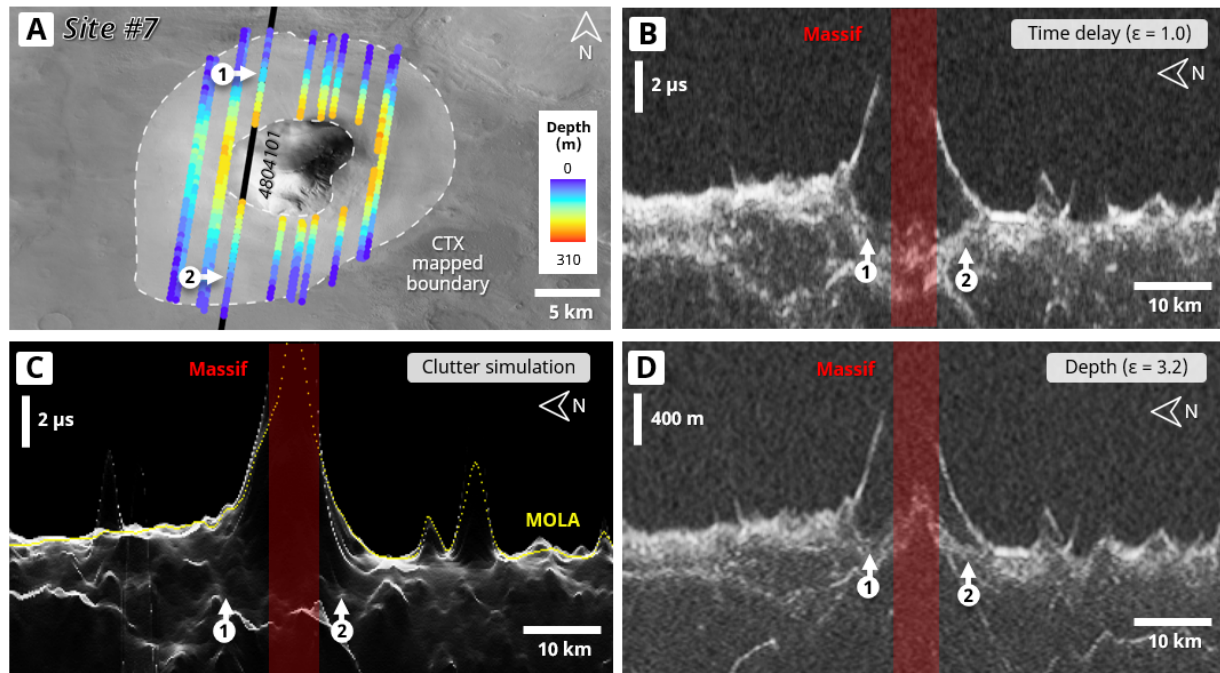


Figure 4.10. Site 7 (A) SHARAD ground track 4804101 highlighted in black with colour scales indicating the locations and depths of subsurface reflectors. (B) Power radargram characteristic of an LDA with a single mound and apron draping the topography. Subsurface reflectors marked as 1 and 2; (C) Clutter simulation of (B) showing the absence of clutter at the same time delay as the reflectors; (D) Depth corrected power radargram ($\epsilon = 3.2$), which shows ice draping the slopes of the massif.

4.3.2.8 North LDA #3 (Site 8)

This LDA is located at 46.45°N , 161.91°E within a composite system due to the presence of a 12-km concentric crater fill to the east (**Figure 4.11**). At the filled crater, we find that clutter inhibits any positive detections; however, subsurface reflectors are present beneath the LDA to the west. The average thickness of this LDA on the western portion is 167 ± 15 m (8.9% uncertainty) with a maximum lag thickness ($d_{1\text{max}}$) of 4 m and a maximum ice thickness ($d_{2\text{max}}$) of 150 m. We calculate a total water-ice volume of $\sim 8.7 \times 10^{10} \text{ m}^3 \pm 7.7 \times 10^9 \text{ m}^3$ excluding the volume of the lag.

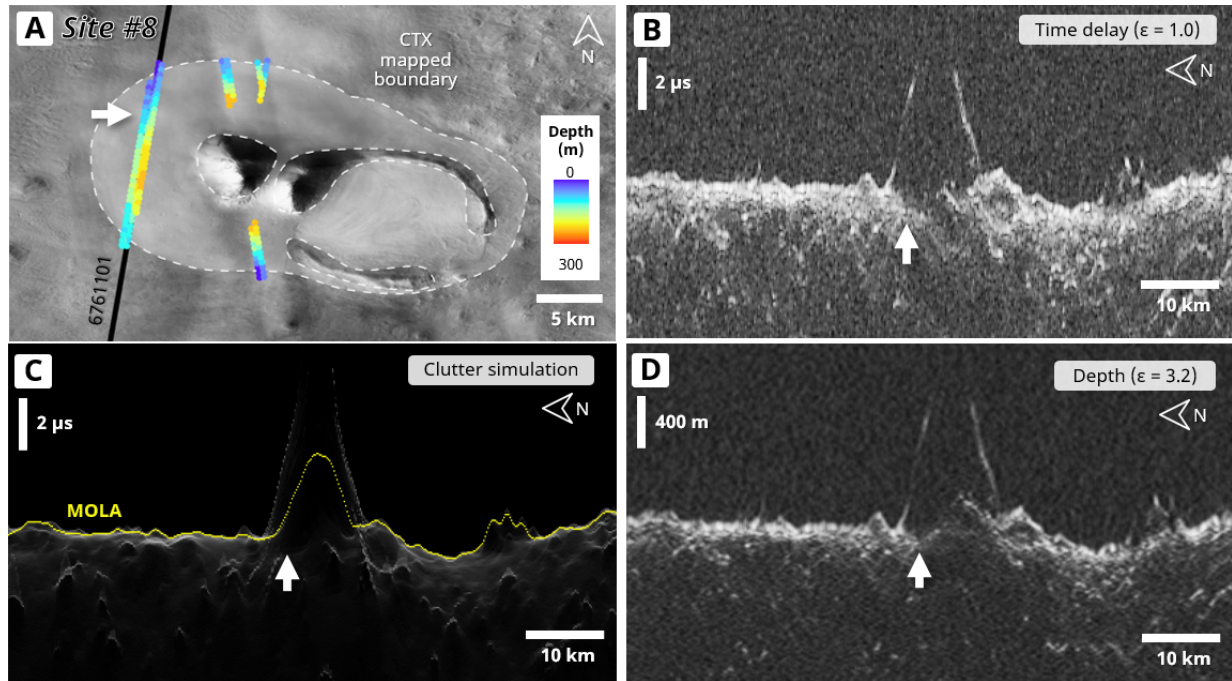


Figure 4.11. Site 8, a composite system with a 12-km concentric crater fill (A) SHARAD ground track 6761101 highlighted in black with the locations and depths of subsurface reflectors in colour. (B) Power radargram of the LDA in panel A. (C) Clutter simulation of (B) showing the general absence of clutter on the apron side to the west; (D) Depth corrected power radargram ($\epsilon' = 3.2$), showing a more representative geometry of the basal reflector.

4.3.2.9 North LDA #4 (Site 9)

Site 9 is located towards the northern boundary of Phlegra Montes, and it is the most northern glacial landsystem detected by SHARAD in the study region (47.82°N , 165.63°E). In Figure 12, we observe an LDA with a very distinct reflector due to its very concave, symmetric basal reflection (**Figure 4.12B**). As the SHARAD orbit track passes through the western portion of the apron, we are able to get a continuous reflector over the massif. This basal reflection when depth corrected (**Figure 4.12D**), shows the convex massif rising above the surrounding plains as the base of the glacier. This geometry is expected for the base of an LDA viewed from this geometry. Like *Site 6* (**Figure 4.9**), this LDA drapes over a massif as it tapers off to the north into the lowlands of the Utopia/Amazonis basin. The average bulk thickness of this LDA is 212 ± 14 m (6.6% uncertainty) with a maximum lag thickness ($d_{1\text{max}}$) of 8 m and a maximum ice thickness ($d_{2\text{max}}$) of 350 m across the different observations (**Table A1**). We calculate a total water-ice volume of $\sim 1.7 \times 10^{11} \text{ m}^3 \pm 4.6 \times 10^{10} \text{ m}^3$, excluding the volume of supraglacial debris.

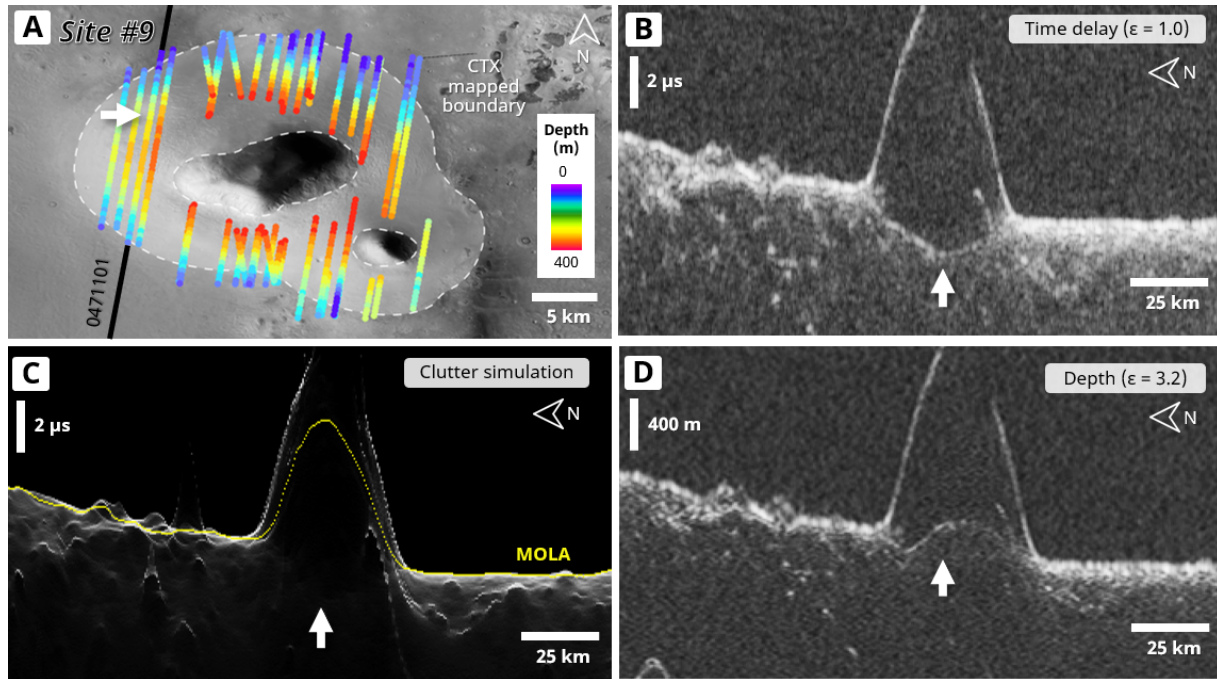


Figure 4.12. Site 9 of the fourth largest LDA in Phlegra Montes (A) SHARAD track 0471101 highlighted in black with the locations and thickness of subsurface reflectors shown in colour. (B) Power radargram of the with subsurface reflectors outlined in the white arrow; (C) Clutter simulation of (B) showing the absence of clutter; (D) Depth corrected power radargram ($\epsilon = 3.2$). In depth correction, the true geometry of the basal reflector is seen. This is what one would expect on the flank of a massif draped with ice.

4.3.3 Dielectric Properties and Loss Tangent

We performed a linear regression to calculate the loss tangents for six out of the eight sites with positive detections, shown in the plots below (**Figure 4.13**). Loss tangent plots in **Figure 4.13** are used to analyze the dielectric properties and infer the composition of the VFF. By comparing the measured loss tangent values with those expected for different materials, we can assess whether the VFF stratigraphic distribution is consistent with clean ice or contains significant impurities, following the approach of Campbell et al., 2008. Some radargrams used in our analysis lack a strong surface return, so some of the sites presented in earlier sections do not have a corresponding loss tangent calculation.

Our results indicate a loss tangent in the bulk interior of five VFFs to have a range of $0.0013 \pm 0.0007 < \tan(\delta) > 0.0026 \pm 0.0009$; except for one VFF in Site 2 that has a $\tan(\delta)$ of 0.007 ± 0.0019 . Loss tangent values between 0.001 and 0.005 are taken as typical for nearly pure water- ice at Martian conditions (Campbell and Morgan, 2018; Plaut et al., 2007; Watters et al., 2007). The occurrence of a higher loss tangent at Site 2 may be due to the heterogeneity of the

material as well as the surface or basal roughness that affects the calculation (Baker and Carter, 2023).

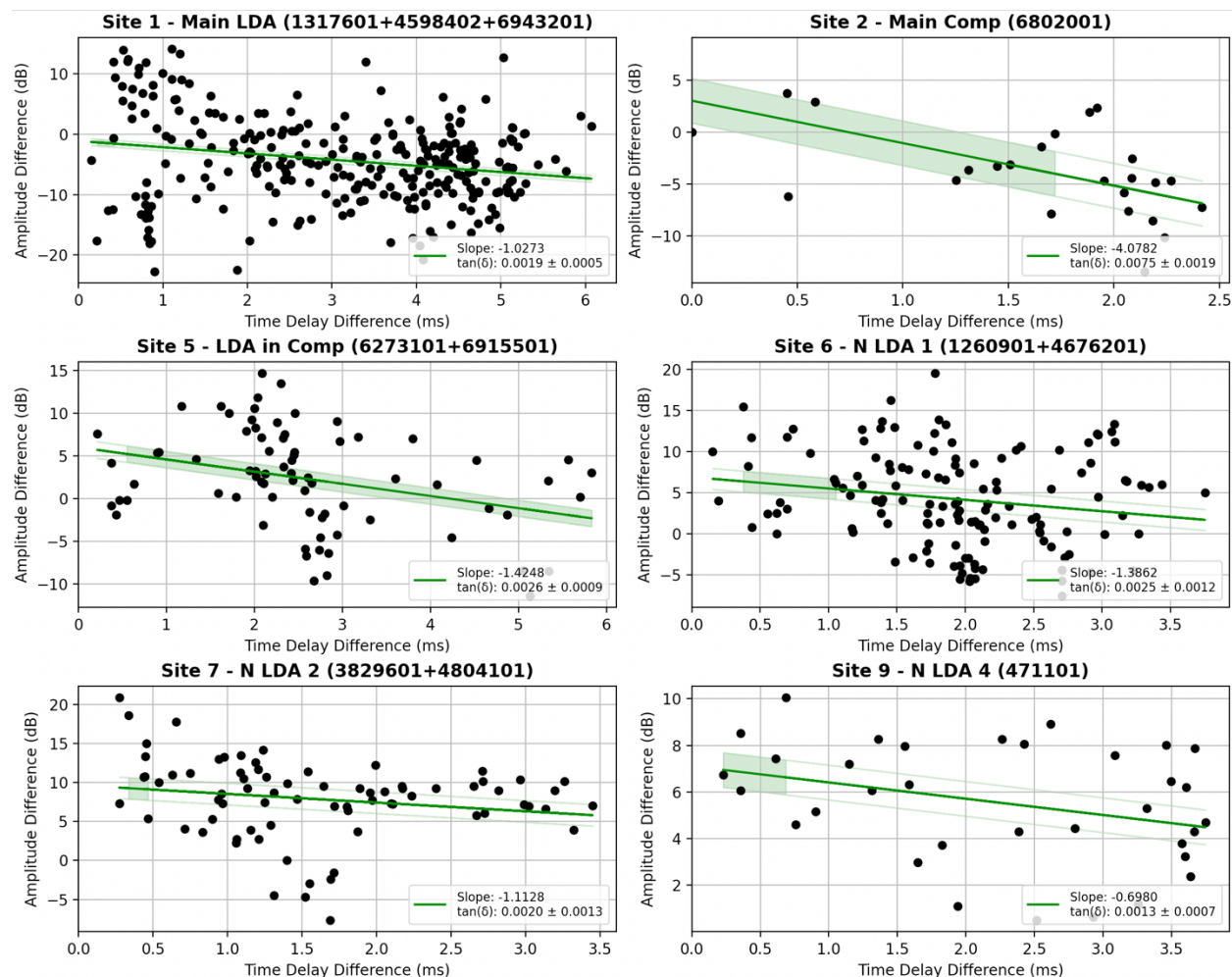


Figure 4.13. Radar attenuation measurements of the six sites with positive detections and clear surface reflections. The relative time delay difference of the surface and the subsurface reflector (x-axis) is plotted against amplitude difference (y-axis). The slope of the best-fit line represents the attenuation rate of the radar signal in the subsurface material. Using Equation 8, we can calculate the loss tangent for each site along with the error propagation of the uncertainty. All sites agree with previous loss tangent values found in other studies of LDAs across the midlatitude regions of Mars (Baker and Carter, 2023; Campbell and Morgan, 2018; Petersen et al., 2018c; Plaut et al., 2007; Watters et al., 2007) with Phlegra Montes loss tangent values ranging from 0.001 and 0.007.

Besides material of differing attenuation (and scattering that lowers the basal reflection strength), the bulk loss tangent is strongly affected by the ratio of the thickness of ice to the thickness of overlying lithic material. For example, a 700-meter-thick glacier with 5 meters of regolith would have a lower loss tangent than a 70-meter-thick glacier with 5 meters of overlying regolith. Thus, the higher value at Site 2, where the average thickness is ~220 meters, may be a combination of four possible explanations: either a surface debris layer composed of very lossy material; non-

negligible volume scattering; the lag-to-ice ratio is high; or the assumption of equal surface and subsurface roughness is false. Internal scatterers such as boulders and or englacial layers are not resolvable by SHARAD. In this calculation, we assume that surface roughness is equivalent to basal roughness.

4.4 Discussion

4.4.1 Mapping Phlegra Montes (Overall System)

The latitudinally widespread glacial geomorphology of Phlegra Montes makes it particularly well-suited for a regional SHARAD study, allowing for a comprehensive assessment of the subsurface structure and ice content of VFFs. Although the mountain range trending from the southwest to the northeast poses an obstacle for clutter mitigation (with nighttime MRO orbits trending from southwest to northeast), eight VFFs were characterized in detail, and we found positive basal reflections with high confidence.

Although there is no clear trend in terms of variations in lag thickness, based on our SHARAD measurements we can see that Site 1 is the largest reservoir of water-ice in Phlegra Montes followed by Site 6, Site 2, Site 9, Site 8, Site 7, Site 5, then Site 3.

Moving from south to north (from Sites 1-9), there is no clear correlation between latitude and the derived loss tangent calculations as most are in the range of 0.001 to 0.003, with only one outside that range. The values found in this study are in agreement with what others found in other glacial systems across Mars' midlatitude (Baker and Carter, 2023; Campbell and Morgan, 2018; Petersen et al., 2018; Plaut et al., 2007; Watters et al., 2007; Steinberg et al., 2025). Our analysis of loss tangent and two-way travel time suggests that the VFFs in Phlegra Montes are primarily composed of water ice.

Null and mixed detections at Phlegra Montes are also present at Sites 3 and 4.

Geomorphologically, visible imagery (CTX, THEMIS) and basemaps guide our SHARAD survey showcasing multiple target VFFs; however, due to the regional topography and orientation of Phlegra Montes, study areas closer to or located on the main mountain ridge are not able to be detected by SHARAD due to the presence of clutter. For example, LDA reflections in composite systems are more prominent to the towards the north because they

follow the path of steeper topography, on pole-facing slopes. Isolated mesas and LDAs towards the west and north of Phlegra Montes are the most easily detected by SHARAD.

In this work, we investigate nine sites with SHARAD; however, there are numerous VFFs that we do not investigate here because of null detections. The absence of detections at visible VFFs may be caused by several factors, some more likely than others. Subsurface composition can affect radar transparency, with higher dust or debris content attenuating the signal before it reaches the base. The depth of the feature may cause excessive attenuation and exceed SHARAD's penetration limits, making basal reflections undetectable (very unlikely for the VFFs here). Additionally, rough or highly heterogeneous surface and subsurface interfaces can scatter radar waves, reducing coherent reflections and further reducing the signal of basal reflections.

Gallagher et al. (2021) found that $\epsilon'_b = 3.6$. While we follow their techniques and agree that $\epsilon' = 3.6$ provides a match using the flat base, this results in unrealistically large lag thicknesses of 35-50 m. Lag thicknesses this large would provide two testable predictions that we find are incompatible with observations: 1) any lag more than ~15 meters thick should result in a resolved reflection internal to the LDA (only true if the contact between debris and ice is not gradational). We do not observe this in any of our sites; 2) thick lags would also show up in the loss tangent analysis (see next section). Because we do not find evidence for either of these predictions, we infer that the debris layer must be thinner than 15 meters.

4.4.2 Significance of terraced glacier systems on Mars

As the first, novel detection of an ice body on terraced topography on Mars, Phlegra Montes signifies a very complex region with potential for studying ice dynamics in high altitude environments on Mars as opposed to other areas with VFFs (i.e. Arcadia Planitia and Deuteronilus Mensae). The presence of terraces in the basal topography of a glacier can provide valuable information about the glacier's past behavior and its current state (Anderson et al., 2018). Similar to topographical variation with rock glaciers on Earth, these terraces are flat or gently sloping step-like features that are often found on the surface of glaciers (**Figure 4.14**). These terraces can form through various processes, and their presence can indicate several important aspects of glacier dynamics, flow, and deformation (Dutrieux et al., 2014; Meng et al., 2023; Petersen et al., 2020; Robl et al., 2015).

While terraces in Phlegra Montes appear to be associated with the composite system where there is a combination of VFFs, it is important to acknowledge that this region is topographically complex. Similar terrace-like features may exist outside of VFFs or in areas where massifs have experienced past glaciation. However, the spatial correlation between terraces and this composite system suggests a potential genetic relationship that warrants further investigation to distinguish VFF-related subsurface modification from other geomorphic processes.

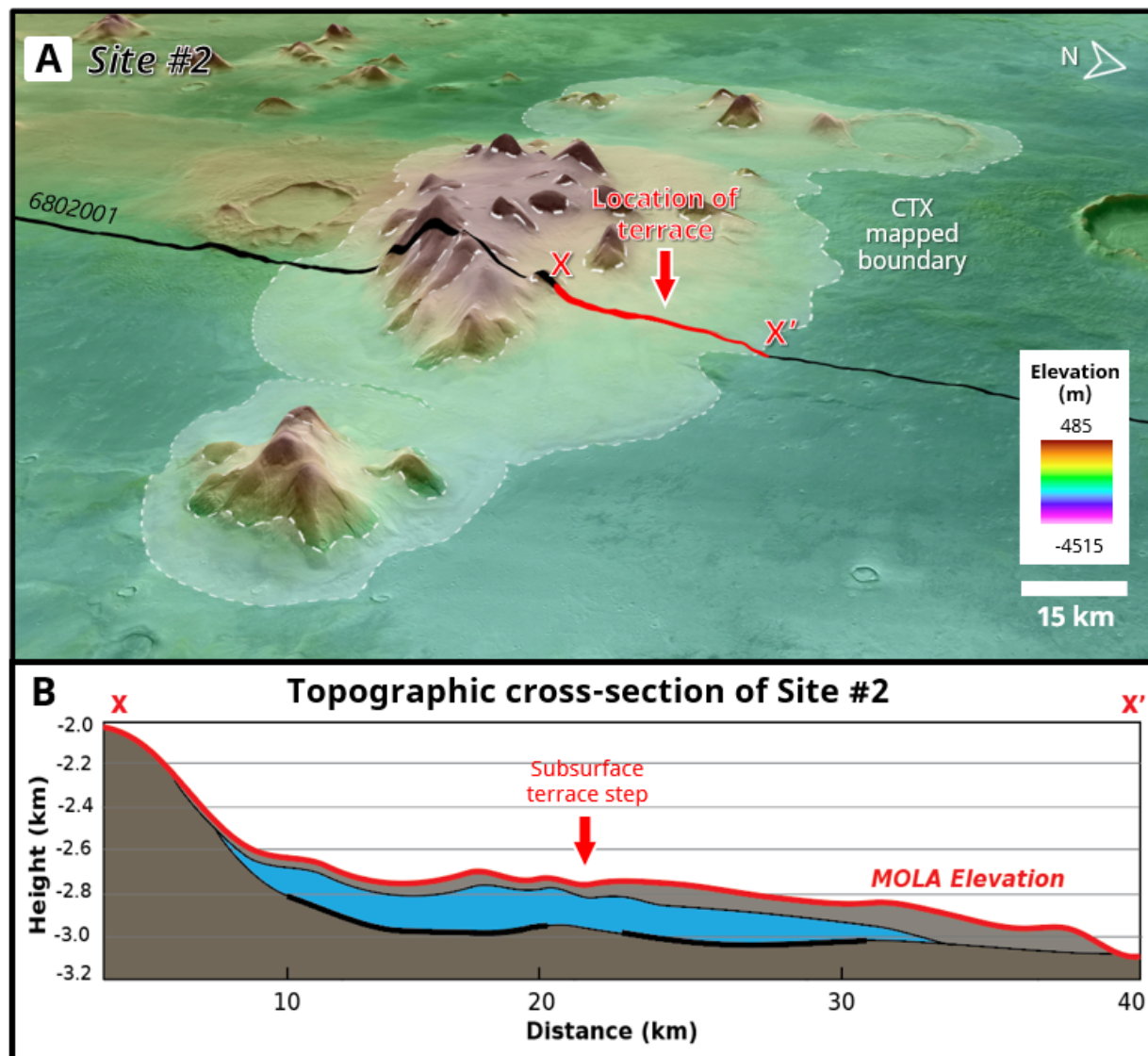


Figure 4.14. (A) A CTX oblique 3D scene draped over MOLA elevation of Site 2: Composite System in Phlegra Montes with a vertical exaggeration of 4. This DEM clearly shows the downhill “flow” of this glacier towards the north with SHARAD orbit 6802001 going over the glacial system. The topographic profile line is highlighted and annotated in red. (B) Topographic schematic of the terraced glacial system with heights derived from MOLA and SHARAD depth corrected radargrams (**Figure 3.5**).

4.4.3 Radar & topographic constraints

Using our analytical solution of a two-layer model (**Equation 7; Table A1 in Appendix**) we determine the lag thickness (d_1) and ice thickness (d_2) from the bulk thickness of the ice body, assuming a specific ice purity in the lower section and a defined dielectric value for the lag (**Table A1 in Appendix**). While there are only two assumptions in this model, they play a crucial role in determining the final thickness values. Variations in ice purity or lag dielectric permittivity could significantly alter the results, highlighting the need for further constraints. Based on our calculations, we find that a bulk dielectric permittivity of $\epsilon' = 3.3$ is appropriate for Site 1. This value agrees well with other studies done in Deuteronilus Mensae (Baker and Carter, 2023; Morgan et al., 2021; Petersen et al., 2018) and across the globe (Steinberg et al. 2025).

In our methods, we have assumed the basal roughness is equivalent to surface roughness to simplify the interpretation of radar data for loss tangent estimates with the lack of direct measurements of the basal properties. Because our loss tangent analysis yielded numbers consistent with nearly pure to pure ice, this supports the assumption that the VFFs at Phlegra Montes primarily consist of ice, with slightly elevated values we attribute to supraglacial debris. However, at one site, the loss tangent analysis yielded significantly higher values. This suggests the presence of additional materials or impurities.

Using equation 7, we are able to estimate lag thickness from the bulk thickness, equation 7 that yields bulk dielectric and assumptions about the dielectric values for the ice and the lag. Unfortunately, we were only able to get one bulk dielectric estimate ($\epsilon'=3.2$) at Site 1. This, plus the arguments based on loss tangent that the ice must be nearly pure and lack of internal reflections leads us to assume that the value applies elsewhere in Phlegra Montes. This provides a lag thickness of ~6 meters at Site 1, and similar values on the order of meters at the other sites. The average percentage of lag to ice at the thickest portion of each VFF is 2%, over a range of ~1.5% to 2.7% calculated directly from the corresponding dielectric permittivity values in **Table A1 in Appendix**.

Geomorphic studies show VFF surface textures indicative of subsurface ice at many locations that we did not analyze here. Larger and deeper VFFs are easier to map using SHARAD due to a

higher rate of positive reflections (away from clutter) while smaller and shallower VFFs in topographically complex regions (i.e. side of valley) in Phlegra Montes were dominated by clutter or were not resolvable with SHARAD. Phlegra Montes is a place with complex topography and rugged, steep slopes. This creates challenges in obtaining clear radar data due to the structure of the landscape resulting in clutter.

4.5 Conclusion

We identified SHARAD reflections at eight VFFs in Phlegra Montes, Mars, seven of which are new glacier radar detections. These VFFs and detections spanned a wide latitudinal range from 30°-50°N in the Martian mid-latitudes. We present the first SHARAD detections of a terraced glacial system (Site 2) and a perched or “hanging” valley glacier (Site 5). We revise previous work and propose a new analytical solution to calculate that $\epsilon' = 3.3 \pm 0.25$ is appropriate for Site 1, slightly under the previously reported 3.6 from Gallagher et al. (2021). Using our newly derived analytical solution for calculating debris cover thickness, we find that the debris at all 8 sites with radar detections is less than 8 meters thick.

A majority of the VFFs we chose to analyze exhibited detectable subsurface reflections; for these, we used a dielectric permittivity of $\epsilon' = 3.3$ to estimate the volume of the LDAs (based on a calculation for Site 1) and calculated loss tangent values ranging from 0.001 – 0.007. Synthesizing our results with the preliminary radar work done by (Gallagher et al., 2021), the findings in this study suggest that debris-covered glaciers in Phlegra Montes house a minimum volume of $1.2 \times 10^{12} \text{ m}^3$ of subsurface water-ice from the detected SHARAD subsurface reflectors. These debris-covered glaciers have a water-ice purity of >95% by volume.

While the density of SHARAD coverage and datasets over Phlegra Montes continues to increase, our understanding of the total ice content within Phlegra Montes remains constrained due to the amount of clutter in the 2D radargrams. Although geomorphologic maps can localize glacial landsystems as a preliminary step of reconnaissance, the application of the 3D imaging technology to the SHARAD observations in this area would clarify the resulting images and greatly reduce clutter, enabling a better assessment of the ice content in this region using a 3D radar image. This site is an important contribution to assessing Phlegra Montes as a strong candidate site for possible In Situ Resource Utilization targets.

CHAPTER 5

5 Distribution and Surface Morphology of Debris-Covered Glaciers and the Latitude Dependent Mantle (LDM) in Phlegra Montes, Mars

5.1 Introduction

5.1.1 Background

Glacial and permafrost ices play a crucial role in shaping landforms on both Earth and Mars, providing insight into planetary processes and histories (Ballantyne, 2018; French, 2017; Matthews et al., 1998; Singleton et al., 2010; Sletten et al., 2003). On Mars, transient ice exposures have prompted hypotheses about substantial H₂O ice reserves in the planet's subpolar latitudes (Dundas et al., 2023, 2021, 2018; Khuller and Christensen, 2021). These deposits, potentially buried under regolith or dust, are shielded from direct atmospheric interactions. Climate models by Forget et al. (2006) and Madeleine et al. (2009) suggest that high-obliquity cycles during past Martian ice ages facilitated the accumulation of significant H₂O ice across the mid-latitudes, extending as equatorward as 30° N and S. Observational evidence supports these models, highlighting a geologically recent, ice-rich formation termed the latitude dependent mantle (LDM). This mantle, present between ~30° and 90° latitudes in both hemispheres, is characterized by a draping morphology that smooths topographic features (**Figure 5.1**; (Conway and Balme, 2014; Kreslavsky and Head, 2002, 2000; Levy et al., 2009). During times of higher obliquity, water-ice from the poles sublimates and is deposited as frost or snow at mid-latitudes, leading to the formation of glaciers and ice-rich landforms (Late Amazonian) (Mellon and Jakosky, 1995). In contrast, low obliquity periods lead to the removal of ice from equatorial areas, while deeper ice deposits remain preserved in the mid-latitudes.

The LDM is estimated to be meters to decameters thick, with dusty ice-rich layers (~30 m thick on pole-facing slopes) with an ice volume exceeding 46% in places (Schon et al., 2009; Conway and Balme, 2014). This deposit has undergone repeated episodes of dissection and re-accumulation (Hollibaugh-Baker and Carter, 2023), contributing to a parallel-layered stratigraphy and a dissected, pitted surface appearance. This deposit often buries much older viscous flow features (VFFs). VFFs, interpreted as debris-covered glaciers, are prominent in

Mars' mid-latitudes and have been extensively studied through geomorphic (e.g., Head et al., 2005), geophysical (e.g., Holt et al., 2008; Plaut et al., 2009; Petersen et al., 2018; Andres and Smith in review), glacial modeling (Fard et al., in review) and climatological (e.g., Madeleine et al., 2009) approaches. VFFs encompass several morphologic subtypes: glacier-like forms (GLFs), lobate debris aprons (LDAs; **Figure 5.1**), lineated valley fills (LVFs), and concentric crater fills (CCFs; e.g., Squyres, 1978, 1979; Hubbard et al., 2011; Souness et al., 2012). They have complex, often curvilinear patterns on their surfaces, which may record histories of ice flow and/or ice deposition patterns/rates. As is the case for glaciers on Earth (Benn and Evans, 2014), patterns on the surfaces of VFFs are likely to reflect complexities in their subsurface structure.

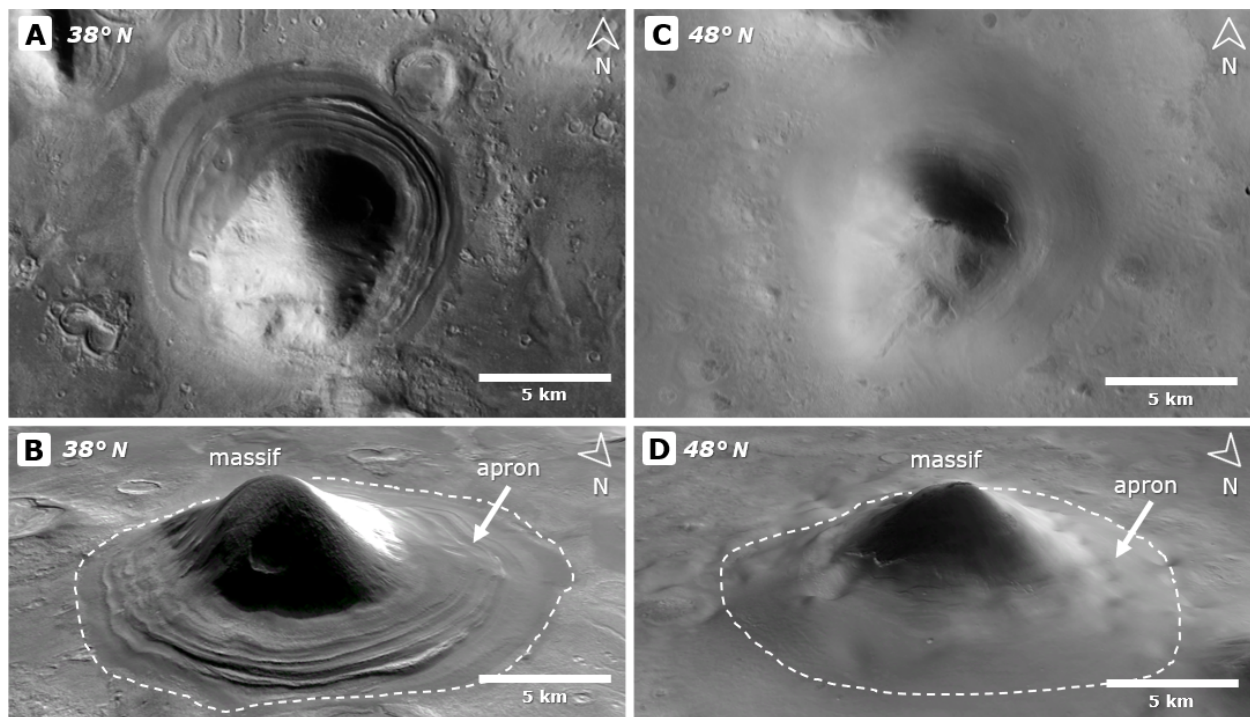


Figure 5.1. Two examples of lobate debris aprons (LDAs) in the Phlegra Montes region at two different latitudes. (A) and (B) show an LDA at 162°E, 38°N with concentric ridges (referred to as lineated terrain in this manuscript) and little to no mantling. (C) and (D) show an LDA at 164°E, 48°N with few distinct surface textures on the CTX scale as it is heavily “mantled”. (A) and (C) are nadir CTX views (Source: CTX controlled global mosaic; v1), while (B) and (D) are oblique CTX views draped over MOLA/HRSC elevation with a vertical exaggeration of 2.

Mid-latitude regions are central to planning for human exploration -- not because ice is most accessible at the surface, as in the polar caps, but because they offer a balance of subsurface ice accessibility and greater solar energy availability. When compared to the polar ice deposits, the Martian mid-latitudes receive greater insolation and are thus “milder” environments for human

exploration. Mid-latitude ice deposits are also important to study because they may represent habitable environments -- they offer protection from surface radiation and may undergo favourable conditions to produce meltwater (Butcher et al., 2016; Cabrol, 2021; Gallagher and Balme, 2015). Thus, VFFs are key targets for future in situ resource utilization (ISRU) of water for sustainable human exploration on Mars, while also serving as valuable records of past climate conditions, which have influenced their formation and preservation (Hoffman et al., 2017; I-MIM MDT, 2022; Morgan et al., 2021).

The origins, chronology, and evolutionary processes of both the LDM and VFFs remain subjects of active debate. Cyclical glaciation on Mars has produced diverse landforms with varying textures and morphologies that may correspond to differences in age, composition, and/or formation processes. Climate models propose ages ranging from <1 Ma to ~500 Ma for both the LDM and VFFs (Forget et al., 2006; Madeleine et al., 2014; Schon et al., 2012). Key research questions persist, such as: **(1)** *Are VFFs multiple generations of LDM deposits?* (Hepburn et al., 2020b; Hubbard et al., 2011).; **(2)** *Are there different types of LDM deposits and can these indicate the nature and/or state of the ice?* (Dickson et al., 2015; Schon et al., 2012). The study of glaciation and climate changes on Mars has provided important insights into the forces behind ice accumulation, leading to the development of glaciated features such as VFFs. Research on recent ice ages (Head et al., 2003; Smith et al., 2016) has been expanded by a more detailed understanding of the climatic drivers over the past 20 million years (Laskar et al., 2002) and improved atmospheric circulation models (Fastook et al., 2011, 2008; Levrard et al., 2007; Madeleine et al., 2009). These advances have refined our knowledge of Mars' climatic and glacial history.

The objective of this study is to investigate latitudinal trends in the geomorphology of debris-covered glaciers in Phlegra Montes, with implications to the climate and subsurface water-ice distribution in the northern mid-latitude regions of Mars. More specifically, we aim to **(1)** create a regional database of VFFs in Phlegra Montes, **(2)** characterize geomorphologic relationships between physical properties and processes that shape them (i.e. glacier confinement, topographic openness, surface facies association), and **(3)** investigate the juxtaposition and possible interaction between the LDM and VFFs.

5.1.2 Regional setting

Phlegra Montes is a mountain range located in the northern mid-latitudes of Mars, centered around 165°E, 40°N, northeast of Elysium Mons (**Figure 5.2**). Stretching from 30°N to 50°N, this region spans a zone across which near-surface ice stability varies, from stable only on pole-facing slopes at low latitudes to ubiquitous at high latitudes. The Phlegra Montes mountain range runs northward and forms part of the Elysium rise complex, and resulted from tectonic uplift during the Early Hesperian to Amazonian period (Fastook et al., 2011; Greeley and Guest, 1987; Hall et al., 1986). The southern edge of Phlegra Montes is positioned near the center of the Elysium rise (28°N, 199°W). In addition, several graben-like structures trend westward and northwestward across the region, with one notable example being a valley system at approximately 40°N that has been filled with debris covered ice (Tanaka, 2005).

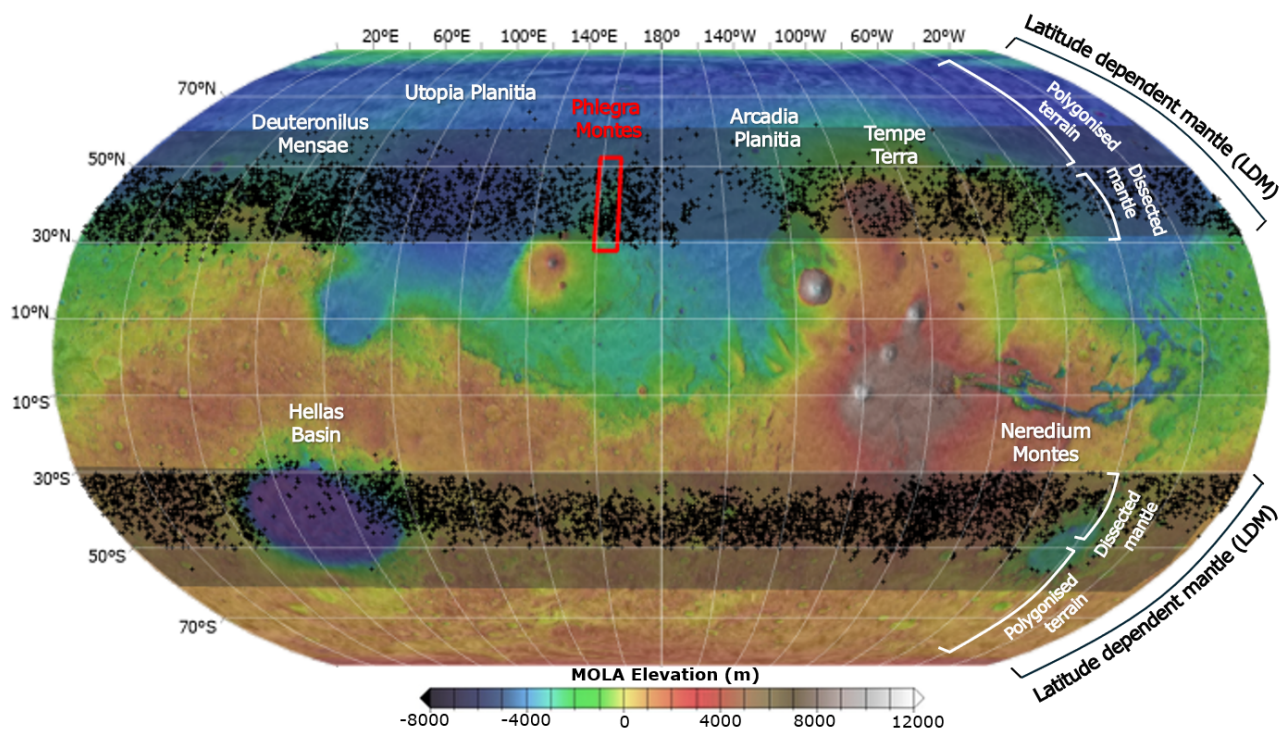


Figure 5.2. Global MOLA map with the distribution of viscous flow features (VFFs) on Mars' mid-latitude region with the Phlegra Montes region study site highlighted with the red boundary. This updated global VFF dataset is a combination of concentric crater fills, lineated valley fills, lobate debris aprons, and glacier-like forms integrated and adapted from (Brough et al., 2019; Levy et al., 2014; Milliken et al., 2003). The mid-latitude region is highlighted in two grey rectangles from 30°-60° in both northern and southern hemispheres, with 30°-50° categorized as "Dissected mantle" terrain and >50° as "Polygonised" terrain (Kreslavsky and Head, 2002). Map data sources: USGS, NASA JPL.

Today, the ice found in Phlegra Montes is a remnant of past climate cycles, which have significantly influenced the region's geomorphology, giving rise to VFFs in the complex topography over time. Phlegra Montes presents an exceptional opportunity to study the cryosphere of Mars across an important latitude range, offering insight into water-ice distribution and its potential for use as an in-situ resource, as proposed by McEwen et al. (2020). Previous research (Dickson et al., 2010; Gallagher et al., 2021; Gallagher and Balme, 2015; Kress et al., 2010; Andres and Smith, in review) has provided evidence supporting the presence of diverse glacial and periglacial landforms indicative of subsurface ice. These include dissected mantle terrain, interpreted as sublimation-modified ice-rich deposits, and polygonised terrain, which forms through thermal contraction cracking of near-surface ice-rich materials (**Figure 5.2**; Kreslavsky and Head, 2002; Levy et al., 2009). Phlegra Montes, on account of its latitudinal range, elevation, and abundance of ice-rich features, serves as a window into Mars' climatic history, water-ice distribution, and the preservation of subsurface ice.

5.2 Methods

5.2.1 Datasets

This research uses imagery from the Context Camera (CTX; Malin et al., 2007; 6 m/px) controlled global mosaic (v1; Dickson et al., 2023), High Resolution Science Experiment (HiRISE; 0.5 m/px; McEwen et al., 2007), and Colour and Stereo Surface Imaging System (CaSSIS; 4.6 m/px; Thomas et al., 2017) as well as topographic data from the combination MOLA-HRSC blended digital elevation model (DEM) dataset (v2; 200 m/pix, Fergason et al., 2018). The MOLA-HRSC DEM was cropped to the extent of our study area from latitudes 28-50° N and longitudes 160-170° E and we derived data products such as slope and aspect using the respective 3D analyst tools in ArcGIS Pro 3.2.2. We also use the MOLA-HRSC DEM to create topographic cross-sections (further described in Section 2.5 of Methods).

Following the methods of Andres and Smith (in review), we also use radar data products from the Shallow Radar instrument (SHARAD; 15 m vertical/px; Seu et al., 2007) and processed through the custom SHARAD processing via the Colorado Shallow Radar Processing System (CO-SHARPS; Putzig et al., 2016). We used SHARAD to investigate radar tracks that intersect

with some of the VFFs in our study to map basal reflectors, which can then be used to derive the thickness of the ice.

5.2.2 VFF Inventory and Mapping

This is the first geomorphologic mapping and inventory analysis of VFFs at Phlegra Montes to fully cover the entire mountain range from 30°-50° latitude. To systematically map this 20° midlatitude stretch covering Phlegra Montes in detail, we divided the 1500 km by 500 km study site boundary into 2° latitude bins using the fishnet tool in ArcGIS Pro. These latitude groups were then labeled resulting in 12 latitude groups in total (including extra latitude groups from the south and north boundaries). The zonal statistics tool was used to extract statistical values for each bin. We mapped VFFs in ArcGIS Pro by digitizing their boundaries to produce polygon shapefiles. We also used the results of mapping efforts done by Brough et al. (2019), Gallagher et al. (2021), Levy et al. (2014), and Milliken et al. (2003) as a guide. To digitize the VFFs, we used high-resolution CTX imagery overlain on MOLA/HRSC topography. The boundaries of the VFFs were determined by identifying the transition from ice-related textures (see Section 2.7) to plains or plateau materials which are often darker, rougher and with more craters. The topographic data was used to assist where boundaries were transitional in nature to identify the breaks in slope where the VFF contacted the plains/plateau. Since we also wanted to highlight isolated LDAs specifically further into the study, we also cropped and extracted the exposed massif heads into separate shapefiles following the same methods. We used zonal statistics to extract elevation and aspect from the MOLA-HRSC DEM within the shapefile boundaries. All imagery mapping and GIS analysis were conducted in ArcGIS Pro 3.2.2, while data visualization and statistical analyses were performed in Python. Using the final shapefiles, we produced a kernel density estimation map of all VFF center points at Phlegra Montes using the Kernel Density tool on ArcGIS Pro 3.2.2. The population field was set to 'None,' and the default kernel radius (bandwidth) was used. Shapefile attribute tables were exported from ArcGIS Pro for further plot creation in Python (i.e. rose diagrams, aspect, confinement; **Sections 5.3.1, 5.3.2, 5.3.3, and 5.3.4**).

5.2.3 Confinement Index

The confinement index (CI) is a measure of the percentage of the VFF that is directly in contact with topography that increases (or goes uphill) away from the boundary and/or in contact with an adjacent VFF. To avoid nomenclature bias (e.g., CCF, LVF, etc.), the confinement index allows us to quantify glacier-phenotypes as a spectrum with two endmembers of confinement. This index ranges from 0 to 1 with values closer to 1 indicating higher degrees of confinement. Low CI values are indicative of VFF on open plains (e.g. LDAs) whereas high CI values are indicative of VFF in highly constrained settings, for example by valley walls (i.e. LVFs), or completely by crater walls (i.e. CCFs). **Equation 9** defines the confinement index with C = length of confined segment of the boundary and UC = length of unconfined segment of the boundary.

$$CI = \frac{C}{UC + C} \quad (9)$$

Confined segments (C) are defined as the sum of the boundary lengths of the glacier that are in contact with any upward sloping topographic boundary (e.g. hillslope, terrace, valley wall, etc.). Unconfined (UC) segments are defined as the sum of the boundary lengths of the glacier where topography slopes away or is flat (i.e. not in contact with adjacent valley, crater walls or uphill boundaries; **Figure 5.3**). To calculate C and UC, the VFF polygon shapefiles were converted to a polyline. Then, the polyline was segmented and categorized into C and UC portions.

The confinement index was calculated for all VFFs mapped in Phlegra Montes. In a second step, we focus on glaciers with lower confinement, as these “simple” systems provide clearer geomorphic relationships between surface features without the complicating influence superposing VFFs (i.e. they are isolated LDAs that remain spatially distinct from other VFFs). This allows for a more direct investigation of latitudinal trends in VFF morphology. Another name for these ‘simple’ systems is LDAs, a term which we use to refer

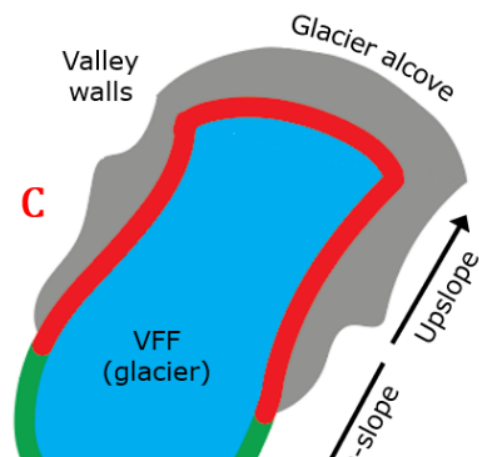


Figure 5.3. Schematic of the Confinement Index parameters. The glacier originates from an alcove restricted within valley walls going outwards onto the unrestricted ground surface. C is highlighted in red as the confined segment, UC is in green as the unconfined segment.

to these features moving forward. Using zonal statistics, we extracted the mean, minimum, and maximum elevations for all VFF shapefiles within each 2° latitude bin from the MOLA-HRSC DEM. We then used Python to calculate the mean confinement for polylines and their corresponding VFFs within each bin (see **Section 4.3.1**).

5.2.4 Centroid Offset and Massif-Apron Ratio

We measured the symmetry of LDAs with two methods by investigating the (1) massif-to-apron area ratio and (2) centroid offset distance as illustrated in **Figure 5.4**. (1) The massif to apron ratio is a measure of the difference in the mapped area of the exposed massif head in comparison to the mapped area of the LDA, minus the massif head. The massif head and the LDA had two designated and separate shapefiles, namely, the massif head polygon and the LDA polygon. The ratio was found by dividing the massif polygon area by the LDA polygon area. The centroid offset (2) is the distance in meters between the center of the LDA polygon and the center of the massif head polygon. The centroid offset distance is also normalized by dividing it by the diameter of the LDA to ensure a dimensionless measurement allowing comparison across LDAs of different sizes. The diameter was measured in ArcGIS Pro by calculating the length across each LDA polygon (using *Calculate Geometry Attributes*, Maximum Length tool). The distance and direction of offset were measured and recorded.

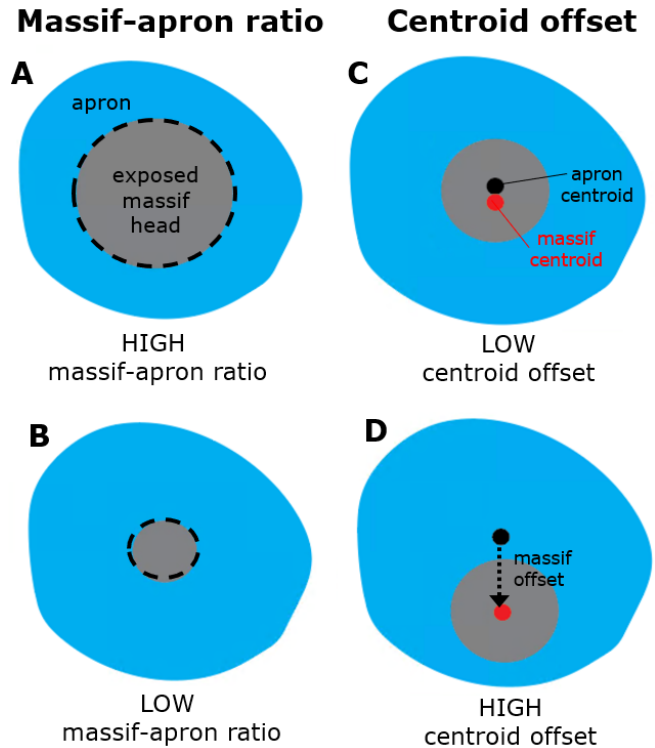


Figure 5.4. Schematics of the massif-apron ratio and centroid offset measurements.(A) An example of a high massif-apron ratio where the area of the massif is larger than the apron (B) An example of a low massif-apron ratio where the area of the massif is smaller than the apron. (C) An example of a baseline centroid shift measurement (of the centroids of the massif and LDA) showing little to no offset. (D) An example of some degree of centroid shift southwards of the massif centroid relative to the center of the apron polygon. Arrow is shown to depict the direction of the offset.

5.2.5 Topography

5.2.5.1 Slope, Aspect and Topographic Profiles

Four initial topographic products and derivatives were extracted from the MOLA/HRSC dataset: (1) S-N topographic profiles, (2) slope, (3) aspect, and (4) topographic openness. (1), (2), (3), and (4) were collected only for LDAs, while (2) and (3) were collected and analysed for all VFFs.

- (1) Topographic transects were manually drawn from the south to north boundaries of each LDA intersecting with the centroid of the massif and LDA polygons. They were then segmented into equidistant sample points with 200 m spacing. This spacing was dictated by the MOLA-HRSC 200 m/px resolution dataset. Points on the transect were then labeled using the mapped the polygon shapefile boundaries as segment categories: *(1) LDA (equator-facing segment), (2) LDA-Massif boundary, (3) Massif, (4) Massif-LDA boundary, (5) LDA (poleward-facing segment)* following clear boundary changes of the shapefiles on the surface of the LDA and massif (**Appendix, Figure A1**).
- (2) Slope analysis was done using the topographic cross-sections to compare variations in slope angles of the exposed massifs and LDAs. We extracted elevation values from the MOLA-HRSC DEM at the position of each of the profile points and used Python to calculate the average slope with a linear fit of each segment category (**Appendix, Figure A1**). This was done in order to find the slope angles of the south- and north-facing portions of the LDAs and the exposed massifs. A separate slope map was also derived using the Slope tool in ArcGIS Pro, which was used as an input to produce an Aspect map.
- (3) Aspect refers to the slope orientation, usually measured in degrees from north. An aspect map was produced using the Aspect tool in ArcGIS Pro. We then used the Zonal Statistics tool to extract circular mean aspect values within VFF polygon shapefiles and used Python to produce plots of poleward-facing slopes versus equator-facing slopes for both exposed massifs and LDAs as well as the aspect of surface textures, described in **Section 5.2.7**.

(4) Topographic openness is a metric that describes the degree of exposure or enclosure of a terrain point or pixel, relative to its surroundings (Yokoyama, 2002). Positive openness represents how exposed, convex, or elevated a location is relative to its surroundings, while negative openness reflects how enclosed or concave a location is. In this study, we calculate negative openness for every pixel of the MOLA/HRSC dataset and used the Zonal Statistics tool to extract mean negative openness to quantify the degree of surface concavity. High negative openness is characteristic of convex landforms such as ridges and hilltops (**Figure 5**), while low negative openness is typical of concave features like valleys and basins. This metric, developed by Yokoma et al. (2002) helps identify and classify landforms by emphasizing the contrast between elevated and depressed areas.

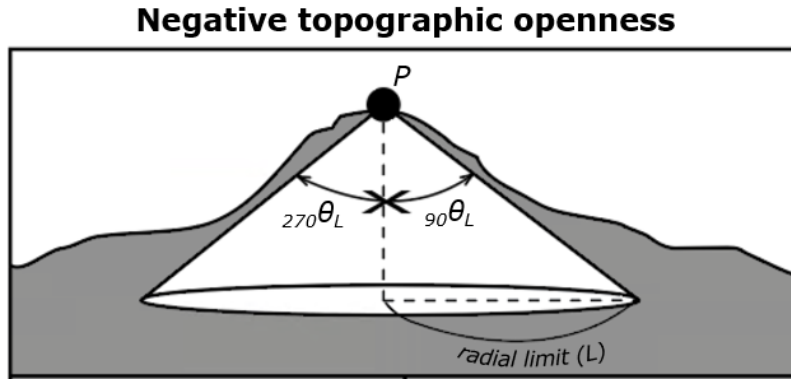


Figure 5.5. A schematic of negative topographic openness. The terrain surface is in grey and L is the radial limit of the calculation for a chosen point on a DEM (Adapted from Yokoma et al., 2002).

5.2.6 SHARAD Mapping

Following the initial detections of Andres and Smith (in review), we analyzed 134 SHARAD radargrams in the survey of single-mound, fully unconfined LDAs. That study derived the equations for bulk permittivity and individual layer permittivity that we use here. They also described methods to define and calculate the loss tangent, which we adopt here.

$$d_b = \frac{ct_t}{2\sqrt{\epsilon_b'}} \quad (10)$$

Where the bulk permittivity (ϵ_b') is calculated from the total two-way time (t_t) and bulk thickness (d_b). To analyze the radargrams, we used geophysical interpretation software (SeisWare™). Targets were selected based on glacier morphology, size, and isolation to ensure SHARAD could effectively detect subsurface features. Using SHARAD data referenced to the aeroid datum, we conducted clutter analysis and depth correction to identify subsurface ice and validate

basal reflectors. A bulk dielectric constant (ϵ_b) of 3.2 for glacial ice (with <15 m of debris) was applied (based on analysis from Andres and Smith (in review) to correct radar signal propagation delays and estimate lag deposit and ice thickness. Radar sidelobes and clutter effects were mitigated to improve confidence in subsurface detections, enabling detailed characterization of the glaciers' thickness. More details about these methods can be found in Andres and Smith (in review).

5.2.7 Identification of Surface Textures

A variety of textures present on the surface of VFFs were mapped using both CTX and HiRISE imagery. We use CTX as the preliminary basemap for mapping as VFFs are on the scale of tens of kilometers in area. CTX is sufficient to detect and map surface textures >18 m in scale. However, surface textures below the scale of CTX were identified in past mid-latitude ice studies (i.e. Levy et al., 2009; Séjourné et al., 2011), pointing to the need for higher resolution. In cases where we find the need, we use HiRISE imagery to identify textures at 1:4000. We use 98 of the ~275 HiRISE images available in the Phlegra Montes study area for texture facies mapping. The remainder were repeat imaging and/or not on VFFs, hence not useful for this study. HiRISE imagery helped to adjust the position of the boundaries between different texture overlaps and to extrapolate textures seen in HiRISE into the CTX mapping scale. The topographic and elevation data from MOLA and HRSC were essential for assessing slope-dependent variations in VFF texture and morphology as well.

In this study, we define five texture categories and assign the textures on the surface of the LDA to one of them. Then we present maps of each based on physical, spatial, and scalar characteristics digitized into polygon shapefiles encompassed in the initial VFF shapefile dataset. These texture facies are adapted from a synthesis of observations from Baker and Carter (2019), Levy et al. (2010), Mangold (2003), Pierce and Crown, (2003) and a preliminary inspection of our mapping area, and we define the five facies classes as: *Lineated*, *Knobby*, *Undulating*, *Smooth*, and *Scalloped* terrains (**Figure 5.6**), further described below. A more detailed texture image gallery and inventory is also illustrated in the **Appendix**.

(A) ***Lineated terrain (>100 m scale)*** was identified using both CTX and HiRISE imagery based on the presence of concentric, cross-slope ridges that extend from or around

massifs (**Figure 5.6A**). These ridges exhibit a consistent concentric orientation and are distinguishable at scales greater than 100 m. Lineated terrain varies in ridge morphology and spacing (~30-150 m). From very sinuous, jagged and angular edges to straight-edged morphologies (**Appendix; Figure A3**) with singular to as many as ~10 ridges. Additionally, lineated terrain often displays alternating bands of darker and lighter contrast, which highlight the ridge structure. HiRISE (where available) was used to define boundaries based on transitions from well-organized, closely spaced ridges to more diffuse, irregular textures. Unfortunately, the topographic expression of concentric ridges is not resolvable in the MOLA-HRSC DEM due to its limited resolution. In some cases, *Lineated* terrain appears as an intervening and isolated texture, so to distinguish between other textures, as a general rule, if the region is >50% dominated by concentric ridges, then the shapefile is categorized into the *Lineated* terrain facies. The recognition of lineated terrain was generally straightforward in regions with one or two superposing textures, but became more difficult in areas or latitudes where the ridges were less distinct with lower contrast and heavier mantling.

(B) ***Knobby terrain (50-100 m scale)*** was identified using CTX and HiRISE imagery based on its rough, highly irregular surface texture. Unlike other facies, knobby terrain lacks an organized structure and

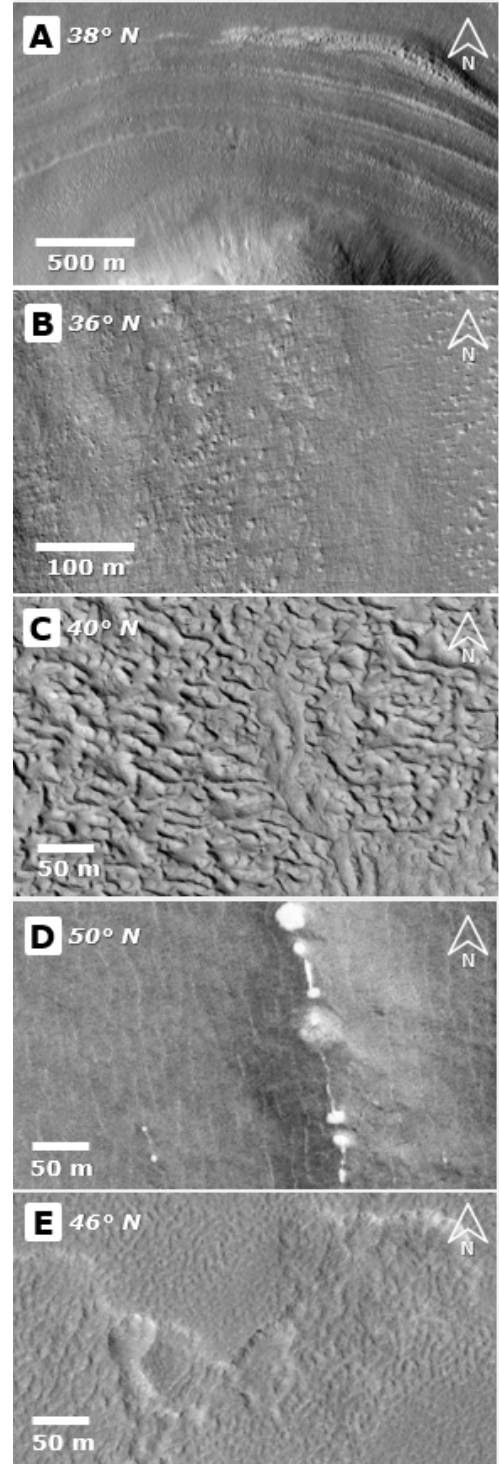


Figure 5.6. Example surface texture facies from each mapped texture category annotated with their respective latitudes. (A) Lineated terrain, (B) Knobby terrain, (C) Undulating terrain, (D) Smooth terrain, (E) Scalloped terrain. (Source: All images are taken from sections of the CTX controlled mosaic v1).

consists of a chaotic assemblage of stippled or mounded landforms (**Figure 5.6B**). These features do not follow a consistent size or shape but appear as a mixture of rounded, elongate, or irregularly shaped knobs, often interspersed with depressions or residual mesas (**Appendix, Figure A7**). This terrain exhibits significant heterogeneity, with variations in texture. At HiRISE scale, in some areas, knobs appear eroded or partially buried by mantling, while others retain steep margins and sharp boundaries (**Appendix, Figure A7**). Mapping was conducted recognizing its consistently rough, stippled texture, particularly near massif peaks or within transitional zones between textures. Some areas display densely packed knobs and pits, while others transition into more dispersed, isolated mounds.

- (C) ***Undulating terrain (10 m scale)*** was identified using CTX and HiRISE imagery based on the presence of crenulated, wave-like surface patterns composed of interconnected sinuous ridges and troughs (**Figure 5.6C**). These features form a distinctive roughness at the decameter scale, with ridge spacing ranging from ~10–50 m. Undulating terrain lacks a single dominant orientation, with ridges and troughs displaying irregular or curvilinear alignments rather than systematic radial or linear patterns. The morphology varies across different regions, with some areas exhibiting tightly packed, low-relief undulations, while others have more widely spaced, pronounced ridges (**Appendix; Figure A4**). Mapping was performed by distinguishing these sinuous, interconnected features from more structured, well-organized ridges (as seen in *Lineated* terrain) or the chaotic stippling of *Knobby* terrain. Boundaries were defined based on clear transitions from undulating patterns to either smoother, featureless regions or more rigidly structured morphologies.
- (D) ***Smooth terrain (<6-m scale; below CTX resolution)*** was classified using both CTX and HiRISE imagery based on the absence of decameter-scale textures. At CTX resolution, smooth terrain appeared as featureless surfaces, while HiRISE imagery revealed polygonal patterns in some areas, clearly distinguishing this terrain from other facies (**Figure 5.6D**). Polygons range from 5-15 m in size with a variety of morphologies (i.e. orthogonal polygons, high-center polygons, etc.; **Appendix, Figure A5**). Mapping relied on recognizing regions where surface roughness was minimal at CTX scale, with boundaries defined based on transitions from visibly textured to smooth surfaces. The identification of smooth terrain was generally more difficult as mantling introduced

subtle, low-contrast textures that blurred the distinction between smooth and “rough” adjacent terrains. HiRISE imagery was crucial in verifying the presence or absence of underlying polygonal structures and ensuring that subtle variations in surface brightness were not misinterpreted as texture changes.

(E) ***Scalloped terrain (<50 m scale)*** was identified using both CTX and HiRISE imagery based on the presence of shallow, rimless scarps or depressions with irregular, curved edges, which form a “scalloped” appearance. These depressions can appear as isolated features or in clusters, coalescing into a network of larger, more complex depressions (**Figure 5.6E**). Mapping was performed by identifying areas where these features exhibited a distinct boundary and regularity in their shape, which set them apart from other surface features. Scallops range from ~60-1000 m in length. HiRISE imagery was essential in confirming the morphological characteristics of scalloped depressions and validating the shape boundaries (i.e. the presence of polygonal patterned ground within or on the surrounding terrain as a criterion for mapping; **Appendix, Figure A6**). To distinguish between *Smooth*, *Undulating*, and *Scalloped*, as a general rule, if the region is >50% dominated by scalloped terrain, then the shapefile is categorized into the *Scalloped* terrain facies.

5.3 Results

5.3.1 Total Mapped VFFs & Confinement Index

We identified a total of 479 VFFs with a latitudinal trend in their confinement. 54% of the total VFFs are located <40° N. Among 479 VFFs identified, 83 are fully unconfined LDAs with a confinement index value of 0 (CI = 0). Of the 83 unconfined LDAs 43% are <40°N and 57% are >40°N. In general, as latitude increases, the altitude of Phlegra Montes VFFs decreases (**Figure 5.7**). The mean confinement of VFFs per latitude bin peaks at 30°-35°N (nearly 75% mean confinement); at 35°-45°N it drops to a mean of 50% confinement, and decreasing linearly to 0-10% confinement north of 45°N. This drop in mean confinement of VFFs is also correlated with the altitude of Phlegra Montes, indicating that there is a strong correlation between glacier confinement and elevation. VFFs located at higher elevations are more confined, whereas VFFs located at lower elevations tend to be more unconfined. This may be caused by the proximity of mountains when the elevation is higher, whereas at lower elevations the reliefs tend to be

isolated mounds. It is also important to note that the frequency of VFFs tend to be lower in both upper and lower latitude extremities (the bulk are between 32°N and 46°N, **Figure 5.7**). In summary, the VFFs and massifs >40°N also appear to be more isolated from the main mountain range as individual LDAs.

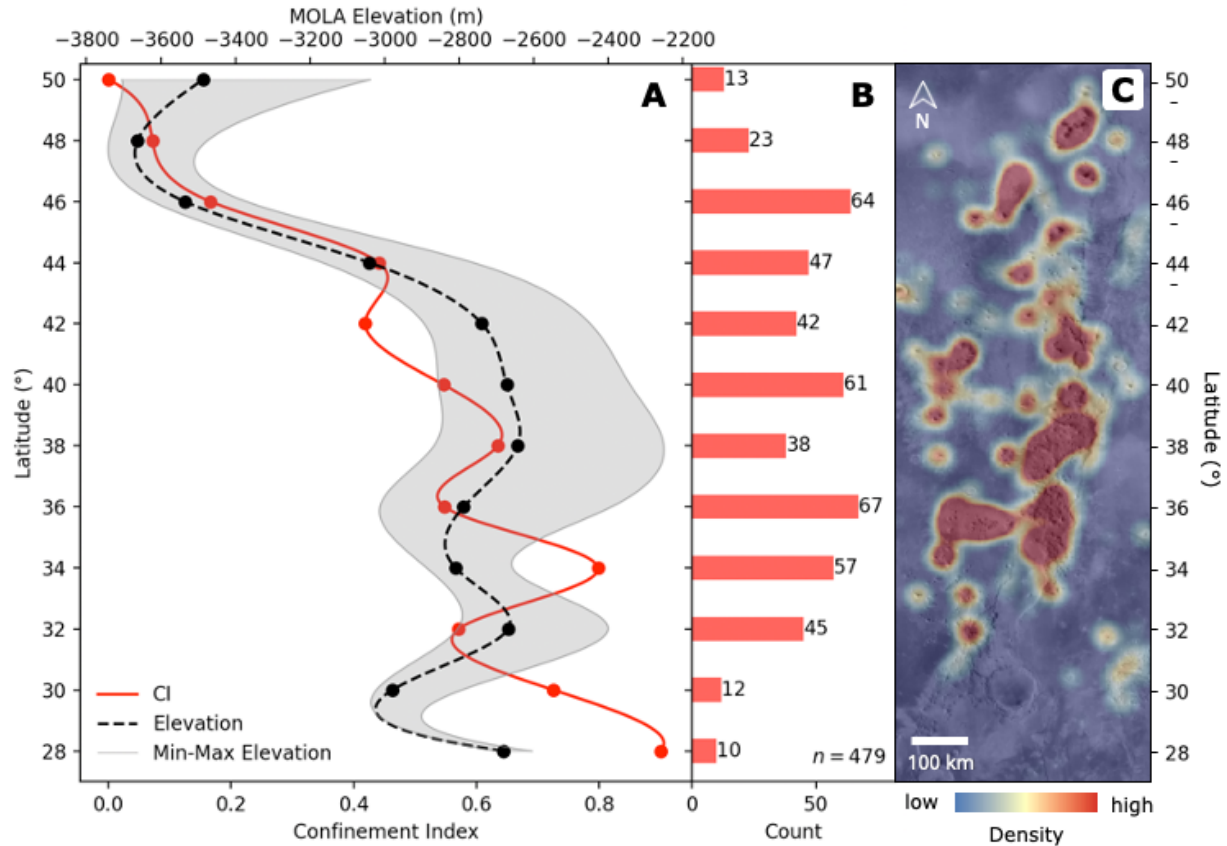


Figure 5.7. (A) Confinement Index of all viscous flow features (VFFs) mapped in Phlegra Montes plotted against MOLA elevation with latitude. The red dots represent the mean confinement for each 2° latitude group and the black dots represent the mean elevation for all sample VFFs in that latitude group. The grey zone represents the minimum and maximum MOLA/HRSC elevation boundaries of the VFFs plotted against latitude (B) The frequency distribution of a total of 479 mapped VFFs in Phlegra Montes per 2° latitude bin. (C) A kernel density map showing the frequency distribution in map view of the VFFs overlain on top of a blended MOLA-HRSC and semi-transparent CTX Mosaic v1.0. (Source: Data are included and available in the supplementary material; Appendix, Table A3)

5.3.2 Massif-apron ratios and Centroid Offset

Focusing on the 83 LDAs, the spatial pattern of exposed massif (bedrock) and surrounding LDA (massif to apron ratio) and distance of the centroid offset can tell us about the asymmetry of these VFF. These measurements are related to the extent of the LDM, ice accumulation and degradation stages of the VFFs. It is important to note that mapping of the LDA boundary shapefiles would not have been possible without the topographic profiles (**Appendix, Figure**

A1) as cross-sections helped determine the transition from LDA morphology to the surrounding flat plains.

In **Figure 5.8A**, we see that the massif to apron area ratio (in red) decreases sharply with latitude, with the lowest latitudes having massifs that are double the area of the VFF itself. Between 35° and 40°N, the ratio of massif and apron areas is 1:1, and >40°N the massifs are nearly completely buried with massif to apron ratios near 0.

In the centroid offset plot (**Figure 5.8B** in black; normalized to the diameter of the LDA), there is an offset of the massif southward for 80 out of 83 or 96% of the LDAs. Additionally, there is also a peak in centroid offset around 40-47°N. This asymmetry can also be seen in plan view or map view in the **Appendix, Figure A2** where all 83 LDAs are displayed per latitude group.

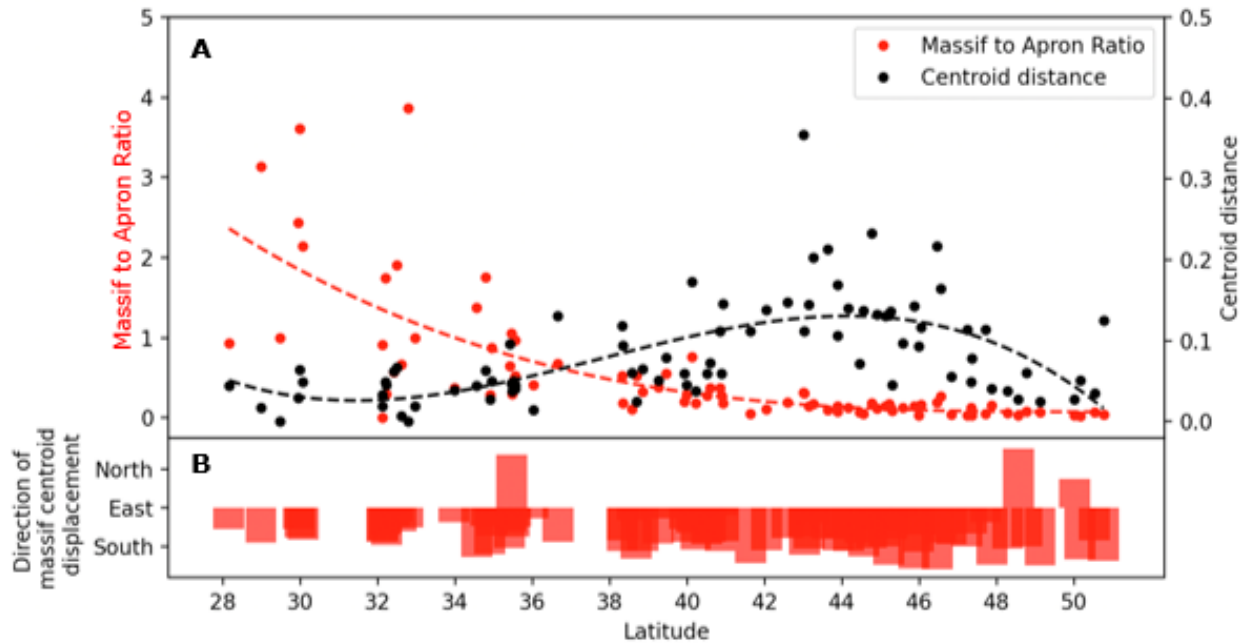


Figure 5.8. Asymmetry of LDAs. (A) Latitudinal trends in massif to apron ratios (in red) and centroid offset distances (in black) with moving averages with a span of 5 points. All centroid shift distances are normalized based on the diameter of the LDA. (B) Direction of the LDA centroid offset from the massif centroid (red bars). Red bars pointing down show southwards offset and red bars pointing up show northwards offset. The y-axis ranges from -90° (North) to 90° (South), making the middle point at 0° (East). The deeper red shading indicates a higher number/density of data points.

5.3.3 Topography and SHARAD

In **Figure 5.9A**, we observe that the negative topographic index (concavity) of the LDAs increases from lower latitudes to higher latitudes. This means that at higher latitudes, the LDA becomes flattened with the massif head slowly being covered by the LDA and the LDA itself

becoming thicker, dominantly on the pole-facing side as shown by the SHARAD data and topographic profiles (**Figure 5.9C**; **Appendix Figure A1**). **Figure 5.9B** shows that the majority of VFF slopes are north- or pole-facing, with approximately 63% (52 out of 83) of LDAs having mean aspects between northwest and northeast. To constrain and estimate the thickness of the ice, we used SHARAD to investigate the depth of the ice to the base of the LDA. Of the 134 SHARAD radargrams mapped across 83 LDAs, only 35 tracks (on 3 LDAs) had positive subsurface detections (**Figure 5.9**; **Table 4.1**). The depth-corrected radargrams show that in agreement with the increasing degree of negative openness to the north, the mean thickness of the ice also increases northwards. The LDA at 48° N has the thickest ice we observe at ~390 m (**Table 5.1**). Although we can get estimations of the total ice depth, SHARAD resolves the basal geometry of the entire VFF system, in this study, we cannot identify the LDM-VFF interface (if any). Thus, surface textures and facies observations are pivotal to further interpreting the origin of the ice underneath using these surface expressions.

Table 5.1. LDAs with positive SHARAD reflectors outlining their mapped ice depths ($\epsilon = 3.2$) and their degrees of negative openness.

LDA ID	SHARAD Track	Latitude	Min ice depth	Max ice depth	Mean ice depth (d_b)	Degree of Negative Openness
50	4804101	43.27°	17 m	277 m	151 m	88.66°
68	1288601	46.01°	23 m	311 m	154 m	88.56°
74	0471101	47.89°	36 m	393 m	223 m	88.90°

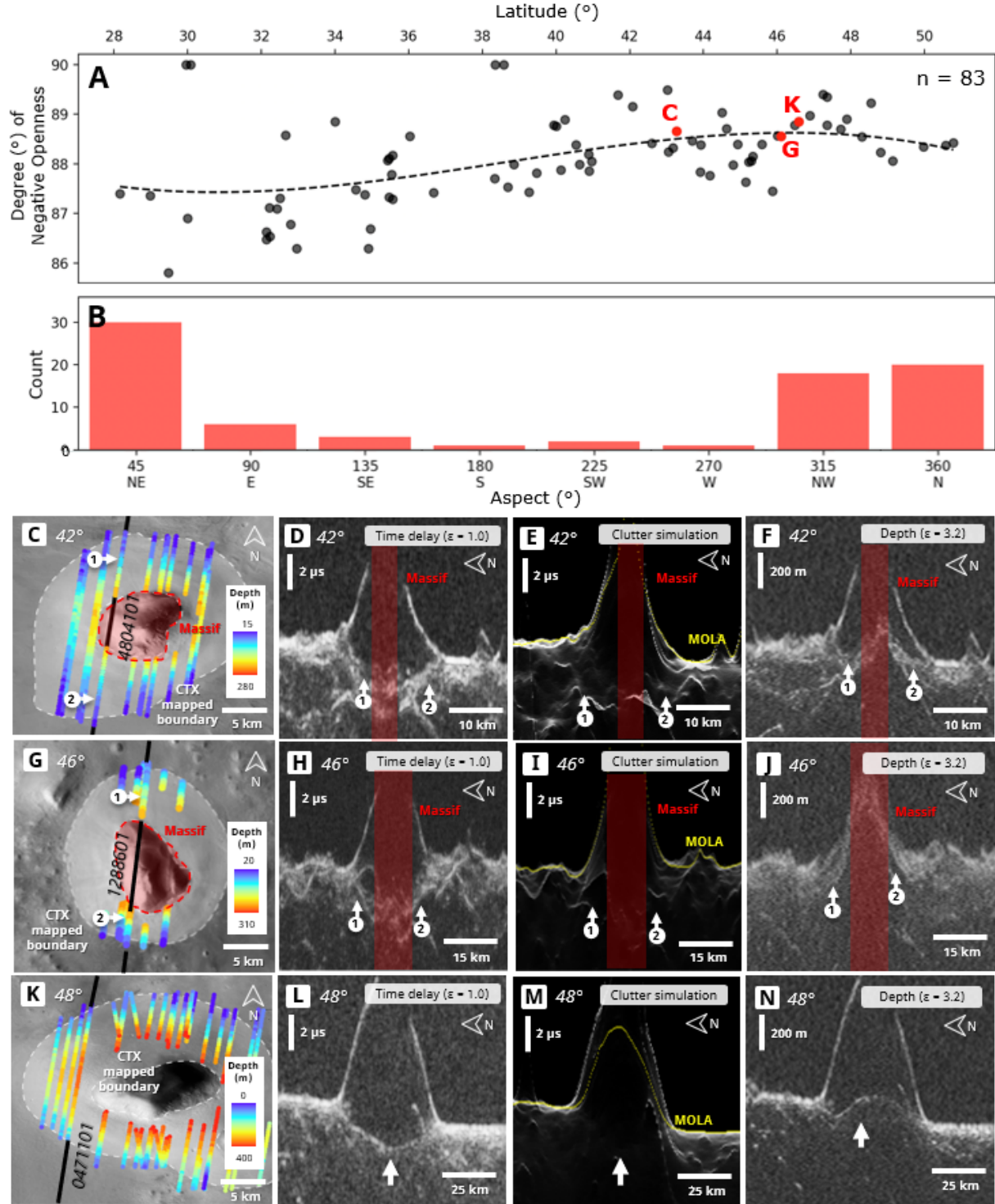


Figure 5.9. (A) Negative Topographic Openness Index of 83 (isolated) mapped LDAs in Phlegra Montes.

Three LDAs with positive SHARAD reflectors highlighted in insets (C), (G), and (K). (B) Mean aspects per polygon of 83 LDAs showing that the VFF slopes are dominantly north-facing or pole-facing. (C), (G), and (K) are SHARAD ground tracks 4804101, 1288601, and 0471101 respectively highlighted in black with colour points indicating the locations and depths of subsurface reflectors. (D), (H), and (L) are power radargrams characteristic of LDAs with a single mound and apron draping the topography. Subsurface reflectors marked in arrows; (E), (I), and

(M) are the clutter simulation showing the absence of clutter at the same time delay as the reflectors; (F), (J), and (N) are depth corrected power radargrams ($\epsilon = 3.2$), which shows ice draping the slopes of the massif.

5.3.4 Surface Texture Facies

5.3.4.1 Spatial Relationship (Latitudinal Trends)

Figure 5.10A shows how the occurrence and density of the five mapped textures vary latitudinally. We note that the *Knobby* and *Lineated* textures are only observed on the surface of LDAs themselves and not outside them, while the *Smooth*, *Undulating*, and *Scalloped* terrains are also found in immediately adjacent plains ~70% of the time from northwards of 42°N. *Knobby* texture is found at all latitudes, while the rest of the four other textures (*Lineated*, *Undulating*, *Smooth*, *Scalloped*) have constrained latitudinal bounds on the surface of LDAs.

(A) ***Lineated* terrain** can be found occurring from ~30°-44° N in 38 out of 83 LDAs with a median location at 36° N. *Lineated* terrain at lower latitudes (<36° N) show lower topographic relief and a more “deflated” appearance compared to *Lineated* terrain in higher latitudes (**Appendix, Figure A3**). At latitudes >40° N, the location of *Lineated* terrain concentric ridges appears to be more distal from the massif rather than directly in contact or adjacent (**Appendix, Figure A2**). The number of concentric ridges observed also decreases towards higher latitudes and they become more isolated, appearing in between other textures. Specifically, 18% of the lineated terrain textures are isolated from the massif and appear within *Knobby*, *Smooth*, and *Undulating* terrains.

(B) ***Knobby* terrain** is ubiquitous in all latitudes with a median location at 40° N. Spatially, this terrain always appears to be in similar locations as *Lineated* terrain at adjacent regions to exposed massif bedrock. If not adjacent to the massif bedrock, it is typically beside *Lineated* terrain. Out of 83 LDAs, 61% have *Knobby* terrain and 39% do not, notably at latitudes >44° N due to the prevalence of other textures. The “knob-like” textures present in *Knobby* terrain are associated with aeolian features, which are also found bordering massif peaks.

(C) ***Undulating* terrain** facies can be found occurring from ~40°-46° N with a median latitude location at 42° N. *Undulating* terrain is the only texture that is found atop VFFs and on terrain adjacent to VFFs, ~20% of the time. As latitude increases (>42° N), the facies appearance and contrast tends to be more subdued. In fact, *Undulating* terrain slowly disappears approaching 48°N (with less surface area; **Appendix**), and the main

texture component becomes the *Smooth* and *Scalloped*. Notably, *Undulating* terrain is 95% of the time found in the presence of *Smooth* terrain. Out of the 83 LDAs, 42 LDAs have *Undulating* terrain present and 95% of *Undulating* terrain appears adjacent to *Smooth* terrain.

- (D) ***Smooth terrain*** can be found from $\sim 32^{\circ}$ - 50° N in 55 out of 83 LDAs with a median location at 44° N. As latitude increases ($>44^{\circ}$ N), the surface area covered with *Smooth* terrain increases (**Figure 4.10; Appendix**). As *Smooth* terrain is present alongside *Undulating* terrain 95% of the time at latitudes where *Undulating* terrain is present, *Smooth* terrain is also accompanied by both *Knobby* and *Lineated* terrain specifically at latitudes $<42^{\circ}$ N on the outskirts and most distal areas of the LDA away from the exposed massif bedrock.
- (E) ***Scalloped terrain*** is constrained to latitudes 44 - 50° N with a median location of 46° N. 25 out of 83 LDAs have *Scalloped* terrain. From 44 - 46° N, *Scalloped* terrain shows more of an isolated spatial arrangement appearing either adjacent to exposed massif heads or on the massif-apron transition boundary. There is a gradual development of *Scalloped* terrain from 46 - 50° N where scallops develop as networks rather than isolated features (**Appendix; Figure A6**). These larger scallop networks ($>46^{\circ}$ N) are found alongside *Smooth* terrain bordering the most distal regions of the apron. Starting at 48° N, the main texture components of all LDAs are *Smooth* and *Scalloped* terrains, which also directly corresponds to the disappearance of *Undulating* terrain at 48° N. Notably, *Scalloped* terrain also displays a concentric pattern, with an average scallop spacing of ~ 400 - 800 m, which is later to compared with the spacing of concentric ridges observed in *Lineated* terrain.

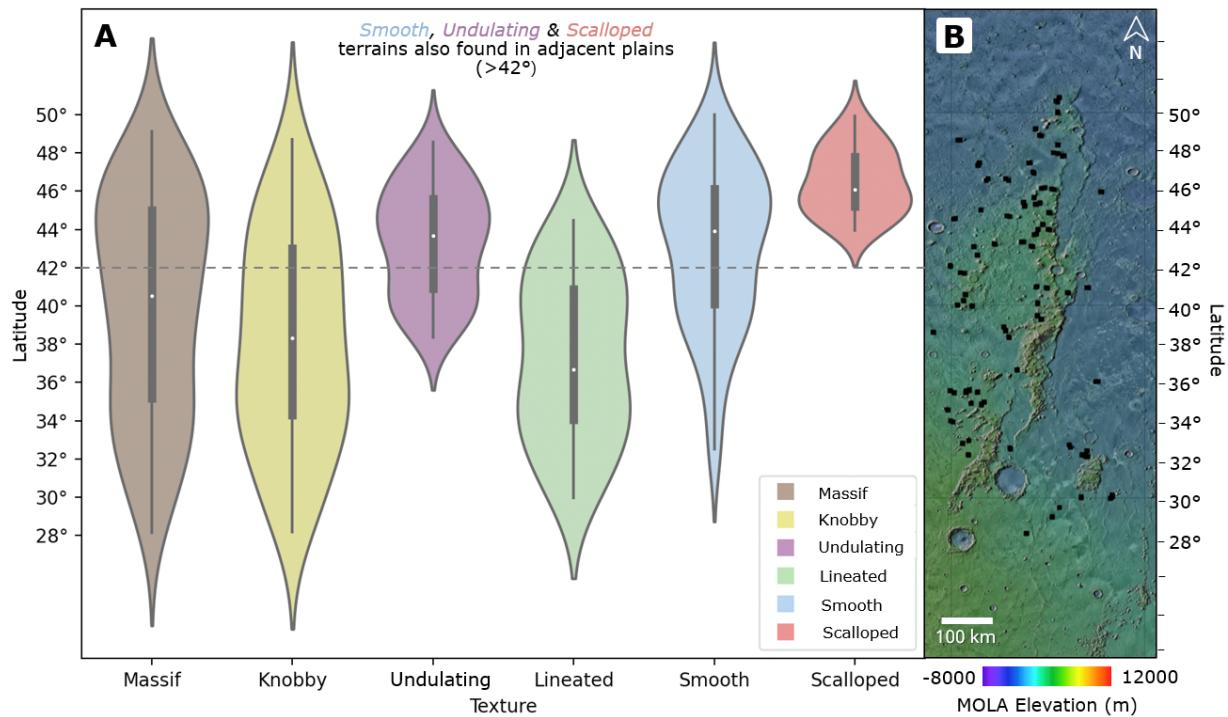


Figure 5.10. (A) Violin Plot of texture distribution and densities over the mapped surfaces of the 83 LDAs. The (x-axis) width of each curve corresponds with the frequency of data points in each latitude region accompanied by a box plot overlay with the narrow vertical rectangle showing the ends of the first and third quartiles with the central dot as the median latitude for each texture group. The massif (brown) is included here for reference but not classified as one of the textures. (B) Location of all 83 LDAs with MOLA/HRSC colourized shaded relief overlain on THEMIS Day IR 100m/px at Phlegra Montes – black dots indicate LDAs.

5.3.4.2 Slope and Aspect

With regards to LDA surface slope, there is no observable asymmetric relationship on pole- and equator-facing LDA slopes, but there is a notable asymmetry between massif bedrock slopes (**Appendix, Figure A1**). As latitude increases (starting >40° N), topographic profiles show that the *Massif* segments on the equator-facing side are 3-5° steeper than the pole-facing side. As shown in Section 3.2 there is also a centroid offset of the exposed massif in the southern section, with more LDA surface area covering northern parts of the massif bedrock, which is also reflected with the shortened *Massif* segments on the pole-facing slopes as it is being covered with more of the LDA.

There is a notable directionality of surface aspects with respect to specific textures (**Figure 5.11**). Our analysis shows that *Knobby* and *Scalloped* terrains primarily occur on southeast-facing slopes, with mean aspect directions of $129.8^\circ \pm 0.8^\circ$ and $167.6^\circ \pm 2.3^\circ$, respectively. *Undulating* and *Lineated* terrains are predominantly found on northeast-facing slopes, with mean

directions of $75.7^\circ \pm 1.0^\circ$ and $97.9^\circ \pm 1.7^\circ$, respectively. *Smooth* terrain appears mostly on northwest-facing slopes, with a mean direction of $301.6^\circ \pm 1.6^\circ$.

The rose diagrams (**Figure 5.11**) indicate that *Lineated* and *Smooth* terrains exhibit strong clustering, with a higher frequency of aspects closely aligned with their mean directions. *Knobby* and *Scalloped* textures show a broader distribution compared to other textures, and the median values (black lines) do not correspond to a maximum in the preferred directions. While the slope and aspect can be important indicators of surface processes on their own, the juxtaposition of these textures relative to their location on the LDA is used to supplement these observations for interpretation

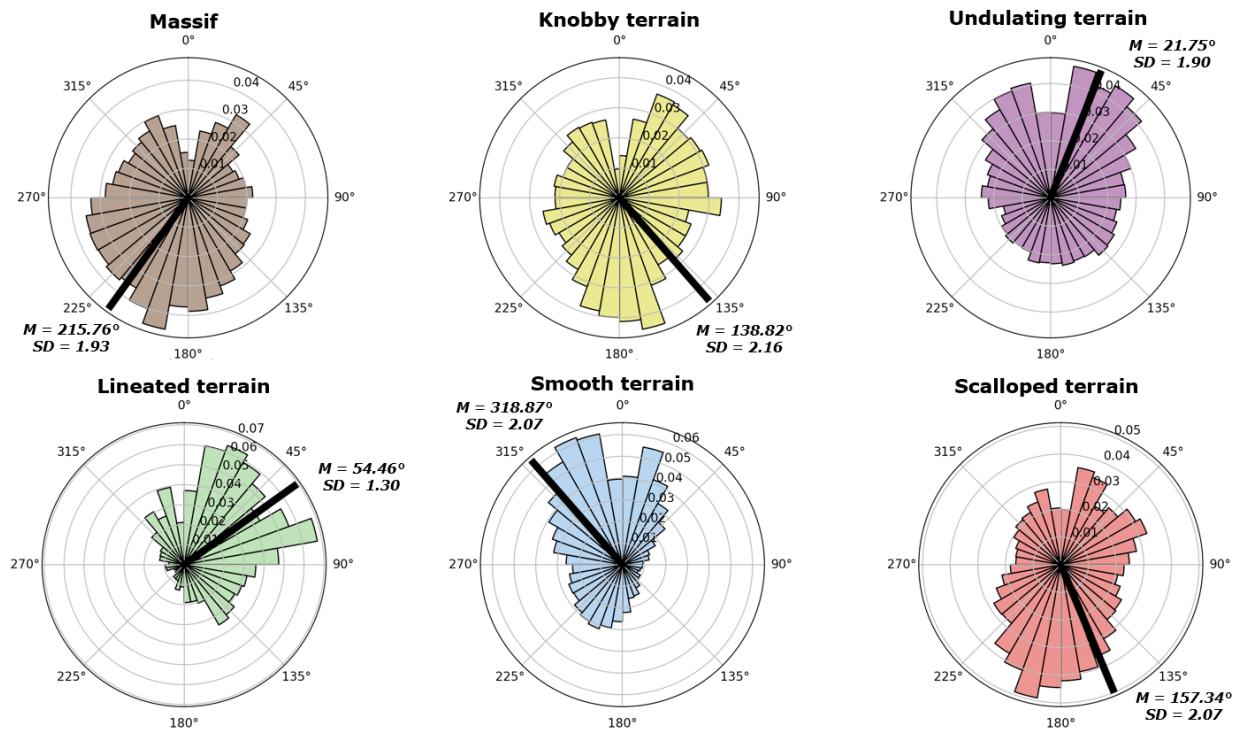


Figure 5.11. Rose diagrams showing aspect directions of each mapped texture on all 83 LDAs with the angular (θ) axis representing directions from $0^\circ/360^\circ$ = North, 90° = East, 180° = South, and 270° = West. The radial-axis shows the surface area covered by each texture. The mean direction is shown with the thick black line and the mean (M) and standard deviation (SD) values labeled in each plot. The massif category was added as a point of reference for the five other textures.

5.3.4.3 Stratigraphic Associations and Boundaries

In general, most texture contacts and topography-slope transitions are gradational and do not express sharp boundaries except when these textures interact with surfaces external to the VFF system (e.g., VFF to plains; **Appendix**).

As we investigated the spatial arrangement of the facies units described in Section 3.4.1, it is also important to investigate if they have any stratigraphic relationships for relative dating and the potential state of the subsurface ice. In map view, the *Lineated* terrain borders the exposed massif heads is at higher altitude than the *Knobby* terrain (**Figures 5.12A and 5.12B**) except starting at latitudes $>42^{\circ}$ N where in 7 out of 33 cases *Lineated* terrain appears as singular isolated ridges within *Smooth* and/or *Undulating* terrain rather than adjacent to *Knobby* terrain. At these latitudes, isolated *Lineated* terrain (**Figure 5.12C**) appears to be covered by the dominant *Smooth* and *Undulating* terrains, which we interpret to represent the gradual burial of *Lineated* terrain beginning at $>42^{\circ}$ N. Notably, *Lineated*, *Smooth* and *Undulating* terrain also have pole-facing aspects which correlates with their juxtaposition on the LDA.

Stepwise, *Smooth* terrain begins to appear at 36° N on the most distal areas of the LDA on both pole- and equator-facing sides (**Figure 5.12B**); furthest away from the *Massif*, *Knobby* and *Lineated* textures. This stratigraphic relationship is true up until 38° N, when *Undulating* terrain appears. *Smooth*-*Undulating* terrains are also textures that appear to have a gradational transition from the LDA extending to the surrounding terrains.

Particularly, where *Undulating* terrain is present, *Smooth* terrain is also present 95% of the time – stratigraphically however, it is inconclusive which of these two units superpose the other. In some cases (38 - 48° N), *Smooth* terrain appears on the outer downslope boundary of the *Undulating* terrain suggesting it might overly the *Smooth* terrain. However, in the same latitude span, the *Smooth* terrain can also appear to overlay or “cover” the *Undulating* terrain (**Figure 5.12C, D**). There is no distinct topographic step or sharp contact between these two textures as they transition into each other.

As the *Undulating* terrain slowly disappears starting at 46° N, the main texture components that dominate $>46^{\circ}$ N are the *Smooth* and *Scalloped* terrains.

Scalloped terrain tends to be adjacent to massifs mainly on south-facing slopes and $\sim 75\%$ of the time (**Figure 5.11**) at slopes of 0° - 10° within *Smooth* terrain. Stratigraphically, *Scalloped* terrain appears as isolated scallops on top of *Smooth* and/or *Undulating* terrain starting at 44° N, and *Undulating* terrain is “replaced” by scallops fully at 48° N. As scallops begin to grow into larger networks, *Scalloped* terrain co-exists with *Smooth* terrain without any clear relationship of

superposition (**Figure 5.12D**). Similar to *Smooth-Undulating* terrains, *Scalloped* terrain can also be found extending from the LDA edges to the surrounding terrain. It is important to note the complete disappearance and burial of *Lineated* terrain from $>42^\circ$ N. We observe that isolated scallops and/or scallops appearing with other textures are often located where *Lineated* terrain is present at lower latitudes, but not visible at higher latitudes. From $>48^\circ$ N, *Scalloped* terrain develops into scallop networks being the main, dominant texture of the LDA.

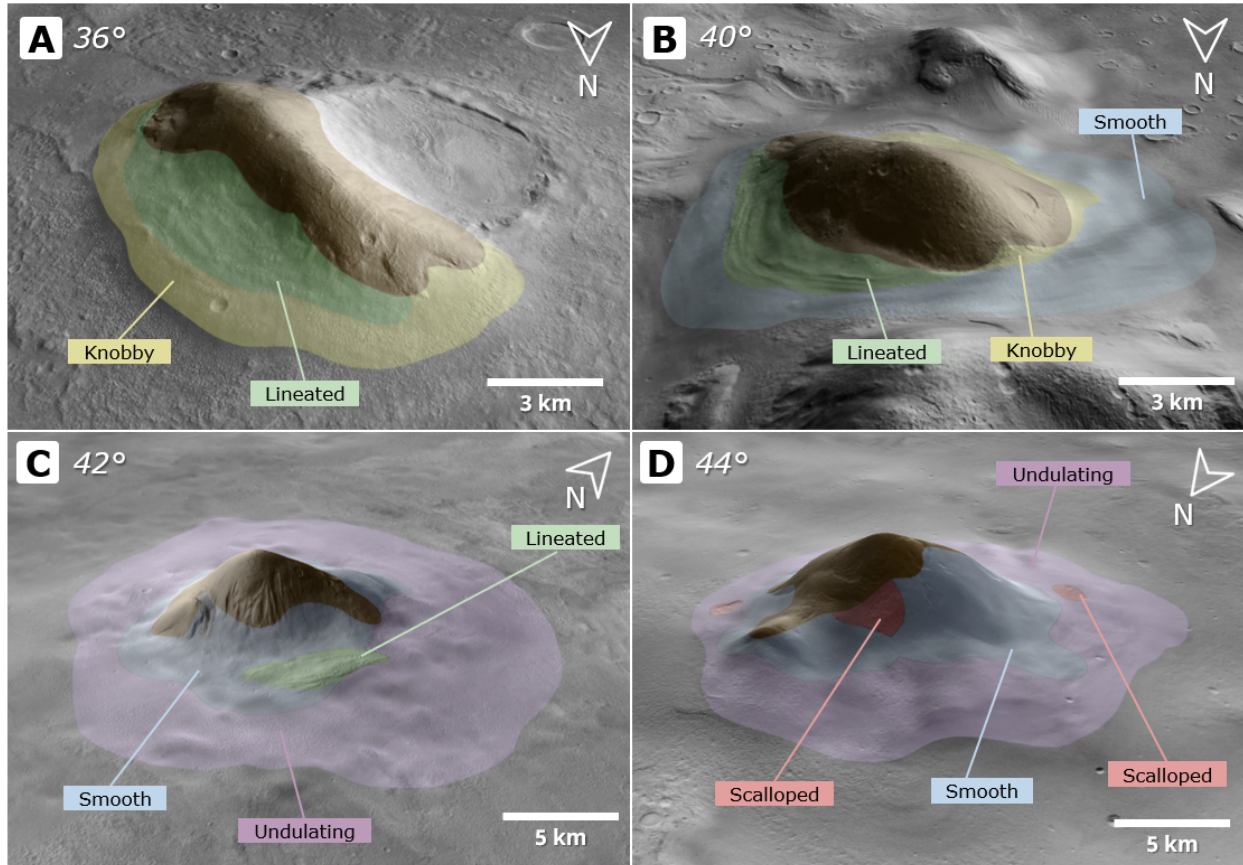


Figure 5.12. Example stratigraphic relationships of textures (Oblique CTX overlay on MOLA/HRSC with a vertical exaggeration of 2). Numbers indicate the order of superposition. (A) An LDA at 36° N with two textures present: Knobby overlain by Lineated terrain adjacent to the massif. (B) An LDA at 40° N with three textures present, noting the appearance of Smooth terrain: Smooth terrain is overlain by Knobby and Lineated terrains. (C) An LDA at 42° N, noting the the gradual burial of Lineated terrain: Undulating terrain is overlain by Smooth terrain. Lineated terrain can be seen intersecting both Smooth and Undulating terrains. (D) An LDA at 44° N, noting the complete disappearance of Lineated terrain: Undulating terrain is overlain by Smooth terrain with some Scalloped terrain appearing in local areas.

5.4 Discussion

5.4.1 Interpretation of surface textures

We interpret *Lineated* terrain to be an expression of compression or debris structures related to deformation of VFFs based on its characteristic parallel ridges and concentric patterns. *Lineated* terrain appears in similar areas as *Knobby* terrain, around and adjacent to the massif. Given the gradational contact between these two textures, we propose that *Lineated* and *Knobby* terrains are both surface textures of the debris cover on VFFs. *Lineated* and *Knobby* terrain may be the result of sublimation and degradation especially at lower latitudes at Phlegra Montes ($<40^\circ$ N), leaving behind an older, more eroded/deflated surface. *Knobby* terrain has also been previously characterized as *Knobby* facies by Pierce and Crown (2003) and *Stippled* texture from Baker and Carter (2019), where they interpret it as degraded mantling deposits. However, we believe that the *Lineated-Knobby* terrains represent the oldest VFF deposits at Phlegra Montes rather than degraded mantle deposits or LDM material (more on **Section 5.4.2**).

We consider *Smooth* and *Undulating* terrains as one unit, because of their ambiguous stratigraphic/spatial relationships. With ~ 3 m imaging data (where shadows on CTX imagery making relief more obvious on north-sides of LDAs), we consider that there could be a systematic under detection of the *Undulating* textures on the south-west sides, which can lead to over detection on the north-east side (shown in **Figure 5.11**; Milliken et al., 2003).

We find that the *Smooth-Undulating* terrains are similar to the “dessicated mantle” or LDM as shown by Kreslavsky and Head (2002). The *Smooth-Undulating* texture variety can be interpreted as a smooth to rough mantle continuum, where the smooth appearance (characterized by polygonal terrain) is the unmodified mantle with higher ice content (as suggested by Levy et al., 2009). We initially thought that the rough/undulating appearance represents areas where ice has sublimated leaving behind lag deposits, however since they appear on north-facing slopes where sublimation is less likely to occur (**Figure 5.11**), we favour the hypothesis that there may be other contributing degradational processes (i.e. thermal contraction and expansion or ice redistribution). *Undulating* terrain morphologies are most similar to *ridge-and-valley* and *sharp-ridge* textures proposed by (Pierce and Crown, 2003), not to be confused with “brain-terrain” (Levy et al., 2009). These surface textures may represent a facies of the LDM .

Lastly, *Scalloped* terrain elsewhere on Mars is often found in areas with high subsurface ice content especially in latitudes $>48^{\circ}\text{N}$ (Mangold, 2011; Séjourné et al., 2011; Ulrich et al., 2010). Our observation that scallops are found on the southern slopes, along with more exposed massifs are likely caused by more insolation that enhances sublimation (Lefort et al., 2010; Zanetti et al., 2010). We interpret *Scalloped* terrain to form in regions of the LDA with very thick ice and potentially high ice purity. This is also systematically shown by SHARAD data in **Section 5.3.3** as the thickest ice deposit of ~ 390 m is found on one of the northern most LDAs dominated with *Scalloped* terrain, while LDAs at lower latitude have thinner ice (**Figure 5.9**). Notably, the spatial pattern of *Scalloped* terrain follows that of *Lineated* terrain -- both appear first as isolated features near massif boundaries, then increase in extent. *Scalloped* terrain also displays a concentric arrangement with spacing between scallops ranging from ~ 400 – 800 m, although larger than the ~ 30 – 150 m spacing observed between concentric ridges in *Lineated* terrain. These similarities in spatial arrangement and morphology suggest that *Scalloped* terrain may represent a surface expression of underlying *Lineated* terrain, possibly as a result of *Lineated* terrain being progressively buried. We attribute *Scalloped* terrain to the most recent degradation stage of icy mantle deposits of the LDM as previously suggested by Dundas et al., (2015), Morgenstern et al. (2007) and Séjourné et al., (2011).

Stratigraphically, since *Scalloped* terrain often occurs superposed on *Smooth-Undulating* terrain (**Appendix, Figure A2**), superposition relationships suggest that *Scalloped* terrain may be the youngest facies unit. However, as noted in **Section 5.3.4.3**, the stratigraphic relationship between these units is not always clear and may vary across sites. We find a similar relationship with *Lineated* and *Knobby* terrains as they occur adjacent to one another with gradational contacts. While we observe *Lineated* terrain at higher elevations than *Knobby* terrain (**Figures 5.12A and 5.12B**), the relative timing between the two remains uncertain due to their overlapping spatial distribution. Overall, as the massif-to-apron ratio decreases latitudinally towards the north, this indicates that the massifs are also systematically being “covered” by LDM-associated textures (i.e. *Smooth-Undulating* and *Scalloped* terrains). These LDM-associated textures are also responsible for the complete burial of *Lineated* terrain ($>42^{\circ}\text{N}$).

We associate this equatorward offset/asymmetry of the massif exposure with the preferential accumulation of *Smooth-Undulating* and *Scalloped* terrains on the pole-facing slope, because ice

tends to be more stable, well-preserved and therefore, thicker on pole-facing slopes (Arnold et al., 2006; Dickson et al., 2012). North-facing slopes receive considerably less solar radiation than south-facing slopes (Arnold et al., 2006; Benn and Evans, 2014). However, ice could be preferentially ablated on the southward-facing slopes, as indicated by the preferential presence of *Scalloped* terrain.

5.4.2 VFF and LDM relationship

There is inconsistency in the literature when distinguishing between the LDM and surface textures associated with debris-covered glaciers (VFFs). Squyres (1978) and Lucchitta (1984) initially described LDM and VFFs as independent, separate systems, but some research suggests they operate as an interconnected system (**Baker and Carter, 2019; Hepburn et al., 2020b**). Variables such as dust availability, subsurface heterogeneity, ice purity/content, topography, and solar insolation may play critical roles in shaping these landforms (Mangold, 2003; Baker and Carter, 2019; Pierce & Crown, 2003).

Our interpretation agrees with previous literature that the LDM is not an integral part of the VFF system. The LDM is a separate component from the VFFs. We believe there is a distinct temporal offset between VFF-associated textures (Knobby and Lineated terrains) and LDM-associated textures (Smooth-Undulating and Scalloped terrains), based on differences in both the topographic profiles (**Appendix, Figure A1**) and stratigraphic relationships (**Figure 5.12**). The LDM can also be seen extending from the LDAs to surrounding terrain (as observed in Smooth-Undulating and Scalloped).

However, this is not to say that there is no interplay between the LDM and VFF. The surface expression of the LDM may very well be modulated or affected by the “deeper” ice from the VFF and vice-versa. We find that Smooth-Undulating and Scalloped textures are degraded states of the ice-rich mantle (LDM), which may play a fundamental role in Mars's glaciation cycle by serving as a widespread, ice-rich source material, that can act as a potential contributor in the formation of VFFs (**Figure 5.13**). Hypothetically, as Mars is currently in a low obliquity period, there should be a net ablation of ice. Given that VFFs have thicker protective, supraglacial lag

compared to the LDM, we hypothesize that it is more likely that the VFFs preserve deeper ice, while the LDM is more susceptible to sublimation due to its thinner lag cover.

Overall, we also note the distinct burial of *Lineated* terrain and disappearance of *Knobby* terrain. We believe that the VFF are buried by the LDM and that the VFF-associated textures and deposits still exist underneath the overlaying *Smooth-Undulating* and *Scalloped* LDM-associated terrains. We infer that the surface expression of *Scalloped* terrain is controlled by the presence of buried *Lineated* terrain. An example is highlighted in **Figure 5.13C and 5.13D** where *Scalloped* terrain is seen to follow a pattern of alternating concentric bands, which is a key characteristic of *Lineated* terrain (as outlined in Methods **Section 5.2**). This leads us to believe the *Scalloped* terrain on top of VFFs can be accentuated by the presence of buried *Lineated* terrain, due to the slope difference in the ridge-valley morphology and radial, cross-slope pattern of *Lineated* terrain.

We use surface textures to further understand key relationships not only between the VFF/glacier and the LDM, but also in relation to the surrounding terrain. More specifically, since the *Smooth-Undulating* and *Scalloped* terrains appear not only on the LDAs but also on the adjacent plains, their distribution suggests that subsurface ice (from the LDM) extends beyond the visible margins of the VFF. The observed bulk asymmetry in surface textures likely reflects the bulk distribution of ice, where ice-related processes influence not only the VFF itself but also the broader landscape, indicating a more extensive ice presence linked to the LDM >40° N (**Figure 5.13B**). Moving northward latitudinally, our hypothesis is that there is a gradational transition from a relict VFF system (shown by “deflated” *Lineated* textures and potentially less ice content) to an environment where the LDM dominates the overall VFF/glacier landsystem (with higher ice content).

5.4.3 Phlegra Montes Glacial Landsystem

These findings directly support our initial objectives of using surface textures and stratigraphy to investigate the evolution of VFFs and their relationship to the LDM at Phlegra Montes. By distinguishing between VFF- and LDM-associated terrains, and highlighting their spatial, morphological, and potential stratigraphic differences, we provide evidence for a transition from an older, debris-covered glacier system to one increasingly dominated by LDM-related

processes. Our Phlegra Montes glacial landsystem model (**Figure 5.13**) suggests a transition from a VFF-dominated landscape at lower latitudes to an LDM-dominated system at higher latitudes. This shift also correlates with decreasing glacier confinement at higher latitudes, where fully unconfined LDAs align with lower-relief regions, highlighting the role of topography and climate in ice stability (**Figure 5.7**):

- (1) **Lower Latitudes (<40°N) – “Relict” VFF System.** VFF-associated textures dominate. *Lineated and Knobby terrains* are well-preserved, indicating past glacial flow and deformation. Massif heads remain largely exposed, with both pole- and equator-facing slopes visible (**Figure 5.8; Appendix, Figure A1**). Ice purity is lower due to prolonged sublimation and lag accumulation, suggesting that these southernmost VFFs have experienced the most ice loss.
- (2) **Mid-Latitudes (40°-48°N) – Transition Zone.** VFFs become increasingly buried by LDM-associated textures. *Smooth-Undulating* terrain appears predominantly on pole-facing slopes, where lower insolation may promote ice preservation. Isolated clusters of *scalloped* terrain appear on south-facing slopes, likely indicating regions of enhanced sublimation. A clear massif-apron asymmetry develops, with pole-facing slopes becoming mantled while equator-facing slopes of the massif head remain exposed.
- (3) **High Latitudes (>48°N) – Mantle-Dominated System.** LDM textures fully dominate, surrounding VFFs beneath with extensive mantle deposits. Ice thickness reaches its maximum (~390 m), and ice purity is highest in these regions. The burial of *Lineated* terrain beneath *Scalloped* and *Smooth-Undulating* textures suggests that scallops preferentially develop along pre-existing VFF-associated textures (**Figure 5.13**).

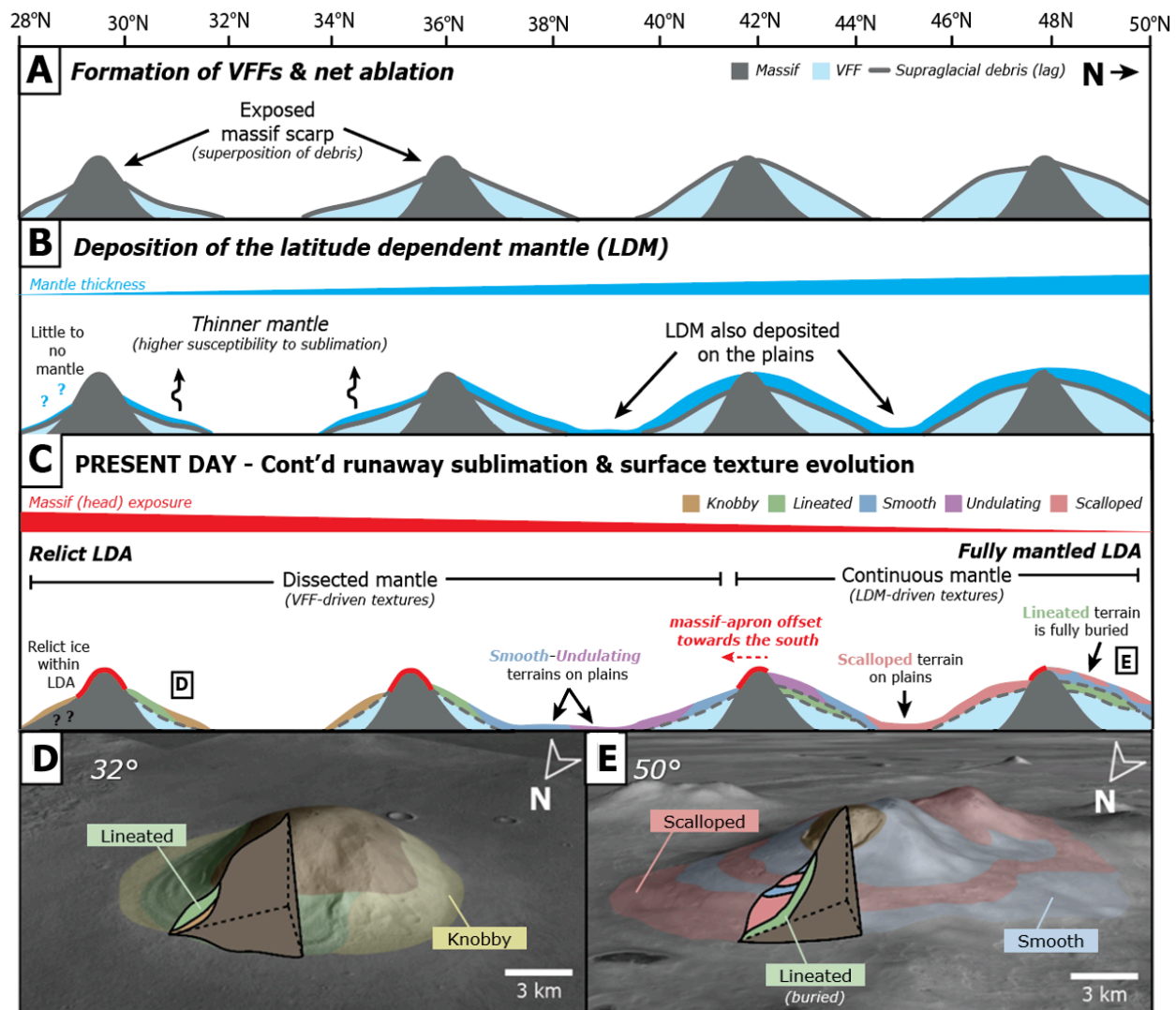


Figure 5.13. Schematic landsystem diagram of the LDM-VFF system as you step through time until the present day Mars. (A) and (B) Emplacement of VFF morphologies and LDM in the Phlegra Montes region at ~300-500 Ma). (C) Present day state of debris-covered glaciers and the “dissected mantle” or LDM as ice continues to degrade from runaway sublimation with surface textures illustrated including their observed aspects and stratigraphic relationships. (D) An inset oblique CTX overlain on MOLA map of an LDA showing the Lineated-Knobby terrain stratigraphic relationship. (E) An inset oblique CTX overlain on MOLA map of an LDA depicting the burial of concentric Lineated terrain by the Scaloped and Smooth terrains. Scalops are occurring following the same radial patterns or lineation shown in (D).

These transitions not only inform interpretations of Mars’ mid-latitude cryosphere but also contribute to broader planetary science goals (Banfield et al., 2020; Planetary Science and Astrobiology Decadal Strategy, 2023). The presence and preservation of buried ice at Phlegra Montes has major implications for long-term climate cycles on Mars, pointing to recurring obliquity-driven phases of accumulation and ablation. From an exploration perspective, the

identification of extensive, potentially well-preserved ice deposits beneath VFFs and LDM textures suggests that this region may be a viable target for ISRU in future exploration missions.

5.5 Conclusion

A regional inventory of debris-covered glaciers in Phlegra Montes identifies 479 VFFs, categorized using a newly developed Confinement Index (CI). The CI quantifies glacier types, ranging from fully confined (CI = 1, e.g., CCFs) to fully unconfined (CI = 0, e.g., LDAs). Our results show that CI values decrease at northern latitudes ($>40^{\circ}\text{N}$), correlating with regional topography, specifically, the presence of fully unconfined, isolated glaciers in low-relief areas. Focusing on the 83 fully unconfined LDAs, we observe a systematic equatorward offset in massif centroid positions (with the furthest offset distance at 44°N) and an increasing LDA coverage compared to the exposure of the massif heads towards the north. This trend suggests that ice preferentially accumulates on pole-facing slopes. SHARAD data from three LDAs further indicate that ice thickness increases with latitude, reaching a maximum of ~ 390 m at 48°N .

We mapped five surface textures in association with the LDAs: *Lineated*, *Knobby*, *Smooth-Undulating*, and *Scalloped* terrains. We interpret *Lineated* ($\sim 30^{\circ}$ - 44°N) and *Knobby* (28° - 44°N) terrains as relict VFF-associated textures. With *Lineated* terrain representing evidence of past VFF movement due to its concentric ridge morphology. Additionally, we conclude that *Smooth-Undulating* ($\sim 40^{\circ}$ - 46°N) and *Scalloped* ($>44^{\circ}\text{N}$) terrains represent the LDM. The *Smooth-Undulating* terrain exhibits a continuum of roughness from polygonal terrain to sinuous ridges and troughs with no distinct stratigraphic relationship, while *Scalloped* terrain represents degradation of the LDM with relatively high ice purity (**Figure 5.12** and **Section 5.1**).

We suggest that the LDM has evolved independently from the VFF system. However, the LDM is modulated by the subsurface ice within the VFF. Shaped by obliquity-driven climate variations, the LDM undergoes episodic deposition and sublimation that create and modify surface features. Over time, interactions between the LDM and underlying glacial ice may lead to the burial or reworking of older VFF-associated terrains. These findings not only improve our understanding of glacial evolution at Phlegra Montes, but also have broader implications. Regions dominated by LDM-associated textures may preserve accessible subsurface ice, making them key targets for ISRU and for the preservation of potential biosignatures, contributing to the ongoing search for past habitability on Mars.

CHAPTER 6

6 Geophysical surveying of ground ice in Tombstone Territorial Park, Yukon: Implications for the International Mars Ice Mapper (I-MIM) Mission

6.1 Introduction

6.1.1 Permafrost and buried ice

Permafrost underlies approximately 50% of Canada's northern territories and Arctic Archipelago, shaping the landscape and influencing ecological and hydrological processes. Permafrost is defined as ground that remains frozen ($<0^{\circ}\text{C}$) for at least two consecutive years, and it houses a diversity of ice-rich landforms within the periglacial domain, including *polygons*, *retrogressive thaw slumps (RTS)*, *pingos*, *palsas*, *thermokarst*, and *buried glacial ice* (Ballantyne, 2018; Dobinski, 2011; French, 2017; Péwé, 1969). Polygonal patterned ground forms due to the thermal contraction of ice-rich ground, creating cracks that fill with water and freeze, often developing into ice-wedge or sand-wedge polygons (Mackay, 1990; Mackay and Burn, 2002). Thaw slumps, a form of permafrost degradation, occur when ground ice melts or sublimates, leading to slope failure and rapid landscape changes that can threaten infrastructure and disrupt ecosystems (Li et al., 2024; Nesterova et al., 2024). Pingos, large ice-cored hills, develop as groundwater is forced upward and freezes under pressure, while palsas are smaller, frost-heaved mounds found in peatlands (Ballantyne, 2018; Kunz and Kneisel, 2021). Thermokarst features, such as depressions, lakes, and alases, form when excess ground ice melts, leading to surface subsidence and altered drainage patterns. Additionally, buried glacial ice, a remnant of past glaciations, can persist beneath permafrost for thousands of years, influencing terrain stability and meltwater availability (Lacelle et al., 2007; Yoshikawa et al., 2006; Andres et al., in review).

Monitoring these ice-rich landforms is critical for understanding permafrost stability, especially as climate change accelerates thaw processes (Smith and Riseborough, 2002; Washburn, 1980). Remote sensing technologies provide powerful tools for detecting, mapping, and tracking permafrost changes across vast and often inaccessible regions. A combination of ground-based, airborne, and satellite remote sensing methods is used to characterize permafrost features (Harris

et al., 2009; Heginbottom, 2002; O'Neill et al., 2019; Ran et al., 2012; Yoshikawa et al., 2006). Ground-based techniques, such as electrical resistivity tomography (ERT), ground-penetrating radar (GPR), and borehole temperature measurements, provide high-resolution data on subsurface ice content and permafrost thermal conditions (Cusicanqui et al., 2023; Supper et al., 2014; Travassos et al., 2008; Watanabe et al., 2013). Airborne remote sensing, including light detection and ranging (LiDAR) and aerial photogrammetry, allows for high-resolution topographic mapping to detect surface deformations related to permafrost thaw (Cunliffe et al., 2019; Kukko et al., 2020). Satellite observations, such as synthetic aperture radar (SAR), optical imagery, and thermal infrared data enable large-scale monitoring of surface changes, subsidence, and thermal anomalies linked to permafrost degradation (Kääb, 2008; Strozzi et al., 2018; Wang et al., 2018). The integration of these remote sensing techniques improves our ability to assess permafrost vulnerability, track landscape evolution, and inform climate adaptation strategies in the rapidly changing Arctic.

6.1.2 Study regions (Site selection)

Our three areas of interest are within Tombstone Territorial Park, Yukon (**Figure 6.1**). More specifically, the study areas are adjacent to Two Moose Lake (64.74°, -138.38°), Chapman Lake (64.79°, -138.36°), and the Chapman Lake Airstrip (64.9°, -138.28; **Table 6.1**) which are located north of the Ogilvie Mountain range. Elevations in the park vary between 535-2345 m.a.s.l. with 95% of the area being above 1000 m.a.s.l. *Site #1: Two Moose Lake* at elevation 1097 m.a.s.l, features well-developed ice-wedge polygons, indicative of permafrost processes and repeated freeze-thaw cycles. *Site #2: Chapman Lake* at elevation 1070 m.a.s.l is characterized by a massive (glacial) ice with a retrogressive thaw slump, exposing relict ice and influencing terrain stability. *Site #3: Airstrip* was a control site at 946 m.a.s.l, as it was flat and easily accessible from the Dempster Highway.

The Ogilvie Mountains, part of the broader Yukon Ranges, have been shaped by extensive glaciation and periglacial activity, with permafrost and ground ice playing a key role in the region's geomorphology (Burn et al., 2015; Frappier and Lacelle, 2021; Grinter et al., 2019; Stockton et al., 2019). Geologically, the range consists of Proterozoic and Paleozoic sedimentary rocks, including limestone, dolostone, and shale, which have been subjected to folding and faulting due to tectonic activity (Amand, 1957; Miall, 1973) . This complex geology has

influenced the region's topography and hydrology, contributing to the formation of permafrost-related landforms (Crites, 2020; Grinter et al., 2019; Kunz and Kneisel, 2021; Lacelle et al., 2007). The study area contains significant permafrost features, including relict ice, segregated ice, and wedge ice, classified by the Canada Ground Ice Map (O'Neill et al., 2019). Understanding these ice formations is crucial for assessing landscape evolution and climate change impacts in the subarctic environment.

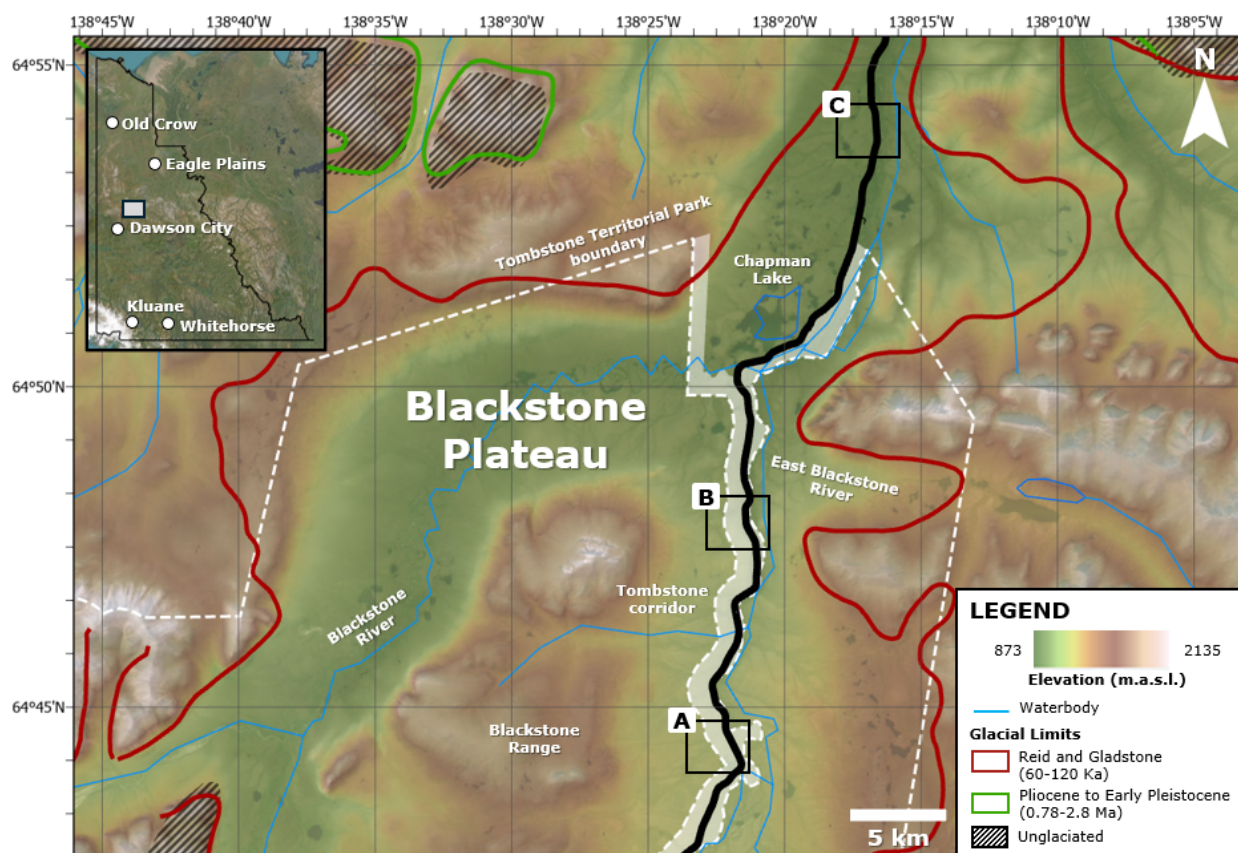


Figure 6.1. Context map of study sites in the Yukon Territory, located within the Blackstone Plateau region of Tombstone Territorial Park. The three sites (A) Two Moose Lake, (B) Chapman Lake Airstrip, and (C) Chapman Lake, are situated along the Dempster Highway corridor. Glacial extents from the Reid and Gladstone advances (60–120 ka) and earlier Pleistocene glaciations (0.78–2.8 Ma) are outlined in red and green, respectively. Inset shows the broader regional context within northern Yukon. Source: YukonGeo and Natural Resources Canada, 2017.

Recent national-scale mapping from the Canada Ground Ice Map (O'Neill et al., 2019) identifies the Ogilvie Mountains and surrounding valleys, including the Tombstone region, as areas with a high probability of ground ice occurrence (**Figure 6.2**). All three study sites lie within zones classified as containing relict ice, segregated ice, and ice wedges, which are consistent with long-term permafrost development and thermal disturbance. These findings are supported by field-

based observations in Tombstone Territorial Park that have documented massive glacial ice, buried ice wedges, and ice-rich sediments (Burn et al., 2015; Camels et al., 2018; Frappier and Lacelle, 2021; Idrees et al., 2015; Lacelle et al., 2009). The exposed glacial ice at Chapman Lake and polygonal ground features at Two Moose Lake align with this classification. Mapping and characterizing these ice types are critical for understanding terrain sensitivity, thaw-induced landscape change, and climate impacts in discontinuous permafrost terrain. This work also complements broader efforts by the Yukon Geological Survey to assess permafrost-related hazards along the Dempster Highway corridor, where similar ice-rich terrain has been the focus of functional permafrost assessments and infrastructure risk mapping (Camels et al., 2018).

6.1.3 Past geophysical studies of buried ice

Geophysical techniques such as ground-penetrating radar (GPR), electrical resistivity tomography (ERT), and seismic reflection have been widely used to investigate subsurface ice in permafrost regions across the Canadian Arctic and globally. In Canada, these methods have revealed relict glacial ice, massive ground ice, and ice wedges in locations such as the Mackenzie Delta, Tuktoyaktuk Coastlands, and Bylot Island (Angelopoulos et al., 2013; Bode et al., 2008; Oldenborger and LeBlanc, 2015; O'Neill et al., 2019). In Svalbard, Alaska, and Siberia, similar approaches have been used to characterize near-surface cryostratigraphy and identify zones of ground ice vulnerability (Arcone et al., 2002; Schennen et al., 2016; Yoshikawa et al., 2006).

In Tombstone Territorial Park, studies by Lacelle et al., 2013, 2009 and Frappier and Lacelle, 2021 have documented the presence of massive relict ice and patterned ground, confirming ice-rich permafrost conditions. These studies complimented and guided our site selection for our present study.

6.1.4 The International Mars Ice Mapper (I-MIM) Mission

These Earth-based studies provide valuable analogues for the detection of buried ice on Mars, where similar periglacial and glacial landforms have been identified. Understanding the preservation, structure, and geophysical signatures of terrestrial ground ice enhances our ability to interpret orbital and in-situ data from Martian mid-latitudes. This research was conducted as part of a technology demonstration for the International Mars Ice Mapper (I-MIM) mission (I-

MIM MDT, 2022), with support from the Canadian Space Agency's Flights and Fieldwork for the Advancement of Science and Technology (FAST) program. The CSA-funded project is entitled *Preparing for Mars Ice Mapper Mission with Airborne SAR, Modeling, and Ground Truth*. The objective of I-MIM is to identify and characterize shallow subsurface ice deposits on Mars using L-band radar sounding from orbit, with particular focus on locating accessible ice resources for future human exploration. The mission aims to detect water-ice to depths of up to 10 meters with a vertical resolution of less than 1 meter and a horizontal resolution on the order of kilometers, sufficient to inform both scientific investigations and in situ resource utilization (ISRU) planning.

Our study contributes to this mission by evaluating how airborne and ground-based radar systems perform in detecting near-surface ice in permafrost terrain on Earth, serving as an analogue environment for Martian mid-latitudes. Specifically, we aimed to (1) use GPR to assess frequency-dependent sensitivity to ice and estimate near-surface ice distribution and volume, (2) apply capacitively-coupled resistivity (CCR) to compare dielectric and electrical responses of frozen ground, and (3) correlate in situ radar measurements with airborne SlimSAR radar sounding data. This integrated approach provides practical insights into the performance of L-band radar systems under variable ice content, layering, and surface roughness, key parameters also expected on Mars. Furthermore, the analogue data acquired here support I-MIM's broader goals of improving radar signal interpretation, refining subsurface modeling approaches, and identifying optimal radar system configurations for Martian exploration.

6.2 Methods

This study employs a suite of geophysical methods to characterize permafrost, ground ice, and subsurface stratigraphy at three Arctic field sites. All spatial data were processed and visualized using ArcGIS Pro, with basemaps and shapefiles sourced from the GeoYukon/Yukon GeoSpatial Data Catalogue (<https://mapservices.gov.yk.ca/GeoYukon/>). The core geophysical techniques used include GPR, airborne radar sounding (SlimSAR), and CCR, which together provide a multi-scale perspective on ice-rich terrain. Each technique exploits distinct physical principles, dielectric contrasts (GPR), radar backscatter and time delay (SlimSAR), and electrical resistivity (CCR) to detect subsurface variability.

GPR enables high-resolution, non-destructive imaging of the subsurface by recording the two-way travel time of electromagnetic (EM) waves reflected at boundaries between layers with contrasting relative permittivity. These dielectric contrasts arise from variations in density, grain size, water content, and mineralogy (**Section 6.2.1**). Two GPR survey collection methods were employed during our field campaign in April 2023: (1) common offset (CO) and (2) common mid-point (CMP) surveys. The CO method will be the focus of this manuscript, detailed in the next section, while the CMP data will be presented in another publication (by another team member, Abigail Lee).

Radar sounding applies similar EM principles but is typically deployed from airborne or orbital platforms. Although generally associated with deeper penetration due to lower frequencies, in this study, SlimSAR operated at a frequency comparable to the 1000 MHz GPR, enabling overlapping depth coverage. Radar sounding provides broader spatial coverage and polarization diversity, making it useful for resolving buried ice and stratigraphic architecture at intermediate depths (Section 6.2.2). Alongside radar sounding, side-looking synthetic aperture radar (SAR) data were collected during the April 2023 flight campaign. This campaign acquired 14 SAR and 20 radar sounding tracks (10 over Two Moose Lake, and 5 each over Chapman Lake and the Airstrip site). The calibration and analysis of SAR data will be presented in a forthcoming publication (by another team member, Ivan Mishev).

CCR is a non-invasive electrical method that estimates subsurface resistivity by transmitting high-frequency AC signals through the ground via capacitive coupling. Resistivity contrasts reflect differences in ice content, porosity, and moisture. CCR is especially suited to frozen terrain where traditional electrodes cannot be deployed effectively (**Section 6.2.3**). *Coring* was conducted at select sites to investigate subsurface profiles (**Section 6.2.4**). These sediment cores provide ground truth for interpreting geophysical data, offering direct observations of ice type, sediment properties, grain size, and stratigraphy. Together, these techniques enable a robust, integrated assessment of near-surface cryostratigraphy and serve as analogue tools for interpreting glacial and periglacial terrain on Mars.

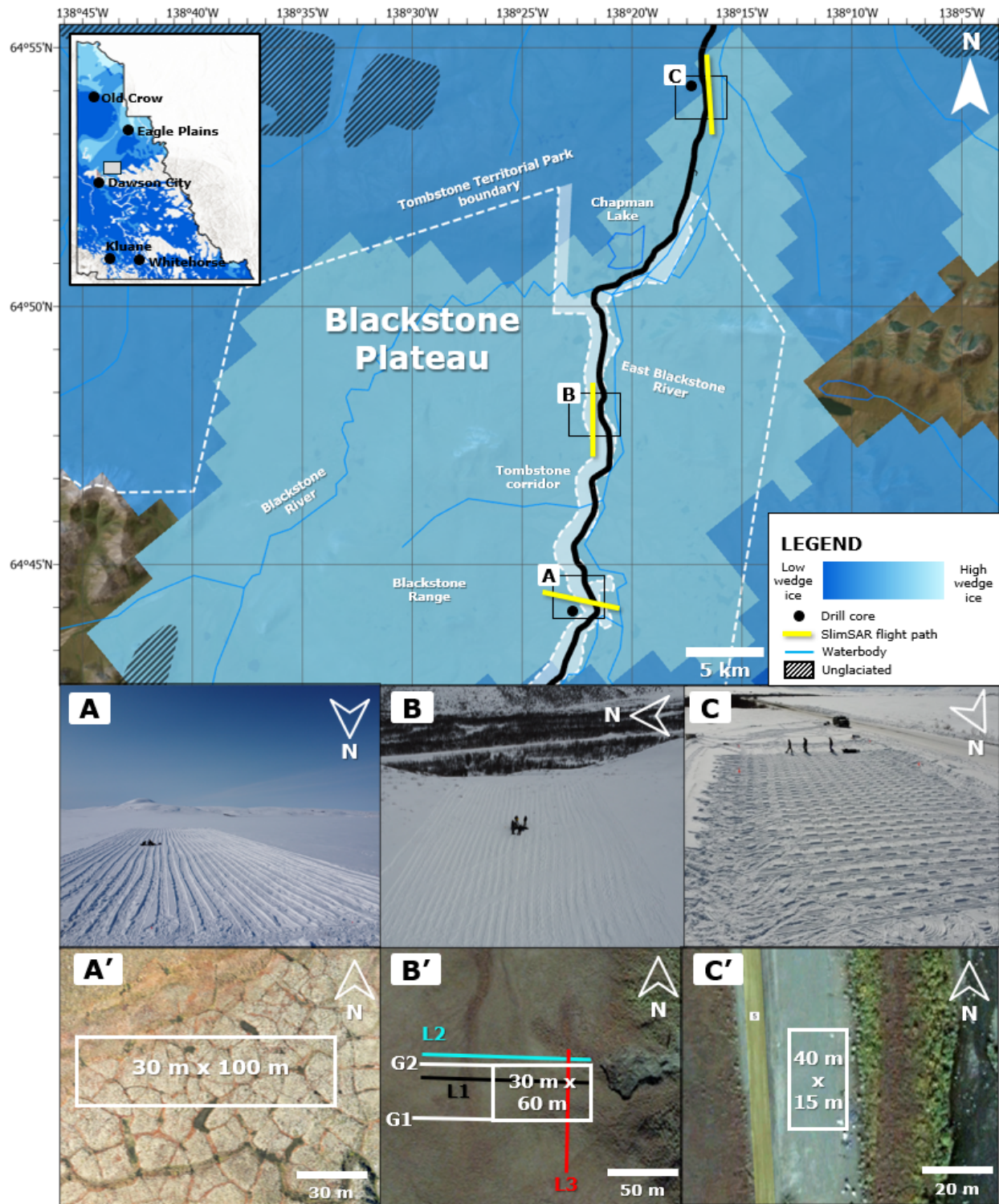


Figure 6.2. Study site locations overlaid on the Canada Permafrost Map showing modelled wedge ice probability (O'Neill et al., 2019). Sites A (Two Moose Lake), B (Chapman Lake), and C (Airstrip) fall within zones of moderate to high wedge ice potential. Yellow lines indicate SlimSAR radar flight paths (April 2023); black dots show drill core locations (June 2023). Insets (A), (B), and (C) show drone imagery of survey grids collected over snow-covered terrain in April 2023, while (A'), (B'), and (C') show orthomosaics of each site highlighting GPR grid

(in white) dimensions and layout. (B') also shows three CCR lines taken in 2009 and 2 additional GPR lines attached to the grid.

Table 6.1. Summary of geophysical surveys conducted at the three field sites in Tombstone Territorial Park: Two Moose Lake, Chapman Lake, and Chapman Lake Airstrip. Methods include GPR, airborne radar sounding (SlimSAR), capacitively-coupled resistivity (CCR), and electrical resistivity tomography (ERT). Snow depth and survey configurations are listed, along with key radar system parameters. Coordinates correspond to the center of each survey area.

		Two Moose Lake Section 3.1	Chapman Lake Section 3.2	Chapman Lake Airstrip Section 3.3
Coordinates		64.74°, -138.37°	64.79°, -138.36°	64.9°, -138.28°
Elevation		1097 m.a.s.l.	1070 m.a.s.l.	946 m.a.s.l.
Degradation type/ Target		Ice-wedge polygons, subsidence	Massive ice, thaw slump/debris flow	Segregated ice
Surveys		GPR Airborne radar	GPR Airborne radar CCR	GPR Airborne radar ERT (GeoYukon) Temp monitoring (GeoYukon)
Snow depth <i>(at the time of surveying)</i>		30-50 cm	30-60 cm	~30 cm
SlimSAR <i>(L-Band radar sounding)</i>	Frequency	1000 MHz		
	PRI	1.73 ms		
	Flight length	3043.80 m	3017.80 m	3409.62 m
	Pulse spacing	5.07 m	5.0 m	5.68 m
GPR	Frequency	200 MHz, 250 MHz, 500 MHz, 1000 MHz		
	Time window	240 ns (200/250 MHz), 37.5 ns (500/1000 MHz)		
	Grid Size	30 m x 100 m	30 m x 60 m 2 lines (100 m each)	15 m x 40 m
	Grid spacing	1.0 m		
	Step size	0.1 m		
Resistivity	CCR N-spacings	-	3 lines (2 E-W, 1 S-N) using dipole- dipole array	-
	CCR N-spacings	-	0.5 to 3.0 m (step 0.5 m)	
	ERT (from Calmels et al. 2017)	-	-	2 lines (Wenner & Dipole- Dipole arrays)

6.2.1 Multi-frequency ground penetrating radar (GPR)

6.2.1.1 Survey and equipment set-up

Our GPR field survey was conducted in April 2023. We used the Sensors & Software PulseEKKO® system with antenna frequencies of 1000 MHz, 500 MHz, 250 MHz, and 200 MHz (**Table 6.2**). These GPR frequencies were “daisy-chained” to run all four frequencies simultaneously in a multi-channel array. With this capability, we used a NIC (Network Interface Controller)-500X, so that receivers can chronologically “listen” to their corresponding

transmitter fire(s). All antennas were mounted on a rigid sled platform and spaced according to manufacturer recommendations to minimize mutual interference: the 1000 MHz, 500 MHz, 250 MHz, and 200 MHz antennas were arranged in a fixed configuration with known offsets (**Figure 6.3**). The sled was towed manually, equipped with an odometer wheel for positioning and a GNSS (Topcon SGR-1) for georeferencing. Power was supplied by a 12V lithium battery, and data acquisition was managed by the NIC-500X controller housed in an action packer along with the receivers.

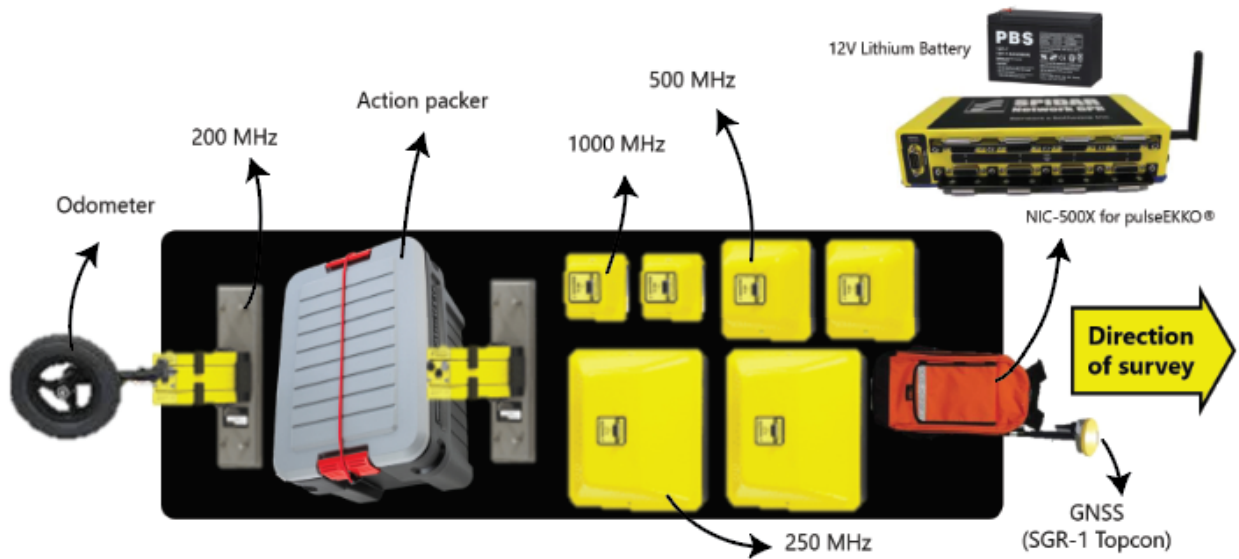


Figure 6.3. GPR sled configuration used in this study, showing the layout of four PulseEKKO® antennas (200, 250, 500, and 1000 MHz) spaced to minimize mutual interference and enable simultaneous multi-frequency data collection. The system includes an odometer for positioning, a GNSS unit (Topcon SGR-1) for georeferencing, a NIC-500X controller, and a 12V lithium battery, all mounted on a rigid sled platform and towed manually in the direction of survey.

At Two Moose Lake, a 60 m × 100 m common-offset survey grid was established using the "zamboni" method, where operators walked in a zigzag pattern with 1 m step intervals. We were able to complete 100% of the 100 m x-axis lines (60 out of 60 lines; long axis) direction of the survey but only completed 60% of 60 m of the y-axis lines (60 out of 100 lines; short axis) due to field time constraints. Chapman Lake was surveyed with the same method over a 100 m × 30 m grid. At the Airstrip site, a smaller 15 m × 40 m grid was used. Data were collected in both N–S and W–E orientations to maximize subsurface coverage. A uniform 1 m line spacing was applied across all sites, which is within or close to the recommended range for the frequencies used: 0.25–0.5 m for 1000 MHz, 0.5–1.0 m for 500 MHz, 1.0–2.0 m for 250 MHz, and 2.0–5.0 m for 200 MHz.

6.2.1.2 GPR data calibration and processing

GPR data were processed using standard routines in EKKO_Project™ software (Sensors & Software Inc.) to improve signal quality and enhance interpretation. Processing steps included time-zero correction, dewow filtering to remove low-frequency noise, background removal to suppress horizontal banding, and gain adjustments to amplify deeper reflections. Velocity calibration was performed using hyperbola fitting in both CO and CMP surveys and reflection profiles to estimate an average wave velocity for depth conversion presented at different sites. Velocity calculation was done using the parabola fitting tool in EKKO_Project™. Where available, topographic corrections were applied using the dGPS on the system. Interpreted radargrams were then analyzed using GPR facies classification based on reflection continuity, geometry, and amplitude patterns.

Table 6.2. GPR frequencies and time windows used with their associated depths of penetration, assuming a velocity of 0.1 m/ns (air or free-space).

Frequency (6 dB bandwidth)	Time window	Antenna Separation (Tx–Rx distance)	Vertical Resolution (target size)	Depth of penetration	Line spacing
200 MHz 100–300 MHz	100 ns	~2.0 m	0.125 m	~10 m	2.0–5.0 m
250 MHz 125–375 MHz	~75 ns	~1.0 m	~0.07 m	5 m	1.0–2.0 m
500 MHz 250–750 MHz	50 ns	~0.5 m	0.05 m	2.5 m	0.5–1.0 m
1000 MHz 500–1500 MHz	25 ns	~0.25 m	0.025 m	1.25 m	0.25–0.5 m

6.2.2 SlimSAR: Radar Sounding

Airborne radar sounding data were acquired using the SlimSAR platform in April 2023, under snow-covered conditions (~50–80 cm of snow cover), over the same survey grids as the ground-based GPR at Chapman Lake, the Airstrip, and Two Moose Lake. SlimSAR is a synthetic aperture radar and radar sounder that transmits linear frequency-modulated (LFM) pulses in the L-band. The system was configured to collect fully polarimetric data with four polarization combinations: HH, HV, VH, and VV. Each polarization was saved in an individual HDF5 (h5) file containing complex-valued radar echoes and associated metadata.

Radargrams were recorded with a pulse repetition interval (PRI) of ~1.73 ms and a sampling rate of 250 MHz. (4 ns of two-way travel time) Each radar trace contains echo samples representing

two-way travel time. GPS position and ECEF-referenced velocities were included to estimate platform motion and along-track distances.

We processed the radar data using a custom Python script, which extracted echo magnitude, cropped the first 600 time samples, and plotted the radargrams in a 2×2 layout (one per polarization). Radargrams were rotated 90° clockwise for display. The x-axis was converted from pulse number to along-track distance using the GPS-derived flight path, which spanned approximately ~3000 m for all sites. The estimated pulse spacing was ~5 m. The y-axis retained raw time sample indices (0–600), labeled from top to bottom, and echo amplitudes were displayed in grayscale. Flight altitudes varied by site: (1) Two Moose Lake: 1466–1477 m, (2) Chapman Lake: 1393–1408 m, and (3) Airstrip: 1316–1326 m. The radar system used a NADIR-looking orientation and flew over the same three GPR grid(s) for direct comparison. Since GPR is typically co-polarized (transmit and receive on the same orientation), we use the HH for direct, structural comparisons.

6.2.3 Capacitively-Coupled Resistivity (CCR)

Capacitively-Coupled Resistivity (CCR) surveys were conducted in 2009 to map ground ice using a Geometrics OhmMapper TR1 system, a non-invasive resistivity method ideal for permafrost terrain where galvanic contact is challenging. The system uses shielded coaxial cables as line antennas, with a 16.5 kHz alternating current capacitively coupled to the ground. As the sled-mounted array was towed across the surface, the transmitted signal passed through the ground and was measured by the receiver dipoles. Apparent resistivity values were calculated based on transmitted current, voltage response, and the geometry of the array, including dipole lengths (5 m) and transmitter-receiver separation distances, or N-spacings. Depth of penetration was approximately 30% of this separation (De Pascale et al., 2008). We employed a bi-directional survey design with N-spacings ranging from 0.5 - 3.0 m (in 0.5 increments), capturing a range of subsurface responses. Data were logged every second at a tow speed of ~0.5 m/s, with positional stakes placed every 20 m along the survey lines to anchor spatial control.

Three CCR survey lines were completed at the Chapman Lake thaw slump, two parallel (E–W) and one perpendicular (S–N), using the dipole-dipole array configuration. Pseudo-sections of apparent resistivity were generated using MagMap2000 software (INC, 2000). Given the high

quality of recorded data, minimal pre-processing was required. Inversions were performed in RES2DINV (Geotomo Software, 2004), producing 2D models of subsurface resistivity. The models allowed identification of high-resistivity zones indicative of frozen ground and potential ice-rich layers (Angelopoulos et al., 2013; De Pascale et al., 2008; Oldenborger and LeBlanc, 2018). Interpretation was guided by site observations of exposed ice-rich sediments along thaw slump headwalls, enabling correlation between resistivity anomalies and known ground ice features.

In addition to CCR surveys, we incorporated existing electrical resistivity tomography (ERT) data collected at the Chapman Lake airstrip (Site DH124) from a previous geohazard assessment study (Camels et al., 2018). Two ERT lines were surveyed at the site using both Wenner and dipole-dipole arrays, one on the field side and one directly on the airstrip. Robust inversion profiles revealed high-resistivity zones interpreted as ice wedges within the upper 3–5 m, underlain by lower-resistivity unfrozen gravels. The surveys indicated that permafrost thickness varied between 8 and 15 m and was highly sensitive to thaw due to the ice wedge complexes near the surface. These archived ERT data supported our interpretation of shallow permafrost distribution and guided integration with our CCR and radar survey results. There is no ERT dataset at Two Moose Lake.

6.2.4 Drill Coring

Manual coring was conducted at Two Moose Lake and the Chapman Lake Airstrip in June 2023 using a gas-powered rotary corer with a 7.5 cm diameter barrel. Cores were extracted to characterize shallow permafrost stratigraphy and provide ground-truth validation for geophysical surveys. At Two Moose Lake, a 1.73 m core was collected from the center of an ice-wedge polygon trough. At the Chapman Lake Airstrip, a 1.41 m core was obtained from a flat, easily accessible site adjacent to the survey grid. Coring was performed until refusal, with permafrost conditions encountered at shallow depths. Core sections were logged in the field for sediment texture, grain size, colour, ice content, and sedimentary structures following the methods of Eyles et al., 1983 and French and Shur, 2010. These samples were used to verify the presence and type of ground ice and stratigraphic layering and will be used in follow-up studies to investigate their dielectric properties.

6.3 Results

6.3.1 Two Moose Lake

6.3.1.1 Multi-frequency GPR

After analyzing ~1050 radargrams, the average EM wave velocity was estimated at 0.162 m/ns ($\epsilon = 3.43$), consistent with values reported for ice-rich permafrost (0.12–0.16 m/ns; Kneisel et al., 2008; Lehmann and Green, 2000). Radargrams in **Figure 6.4** reveal two prominent ice-rich targets located at approximately 11 m and 20 m along the profile, visible as strong reflectors across all frequencies. These anomalies correspond with polygon troughs visible at the surface and coincide with areas of seasonal pooling, supporting their interpretation as ice wedges (**Figure 6.2A'** and **Figure 6.5**).

All radargrams were cropped to 3 m depth (for comparability), but the 500 and 1000 MHz have up to 10 m penetration depth. The 1000 MHz profile resolves fine-scale layering and stratigraphic heterogeneity in the upper meter, though signal attenuation reduces clarity beyond 1 m. In contrast, the 500 MHz and 250 MHz frequencies balance resolution and penetration, capturing wedge-like structures that appear as hyperbolic and dipping reflectors at depths >1.5 m. The 200 MHz radargram displays broader subsurface patterns but with reduced resolution near the surface.

Depth-sliced amplitude maps (**Figure 6.5**) further support the interpretation of wedge ice. Slices were generated at 0.1 m intervals (with lateral 50% overlap), truncated at 1.8 m depth to highlight the ice wedge interface around ~0.7 m. High-amplitude, curvilinear features trace polygonal patterns across slices from 0.8–1.6 m depth, with some wedge-like features extending beyond 2 m. While most ice wedges appear to initiate near the 0.7 m depth horizon, others begin as great delay, suggesting variable geometries and possibly two depths of burial since formation.

These results confirm the presence of multiple generations of well-developed wedge structures beneath Two Moose Lake and hint at stratigraphic complexity in their growth history.

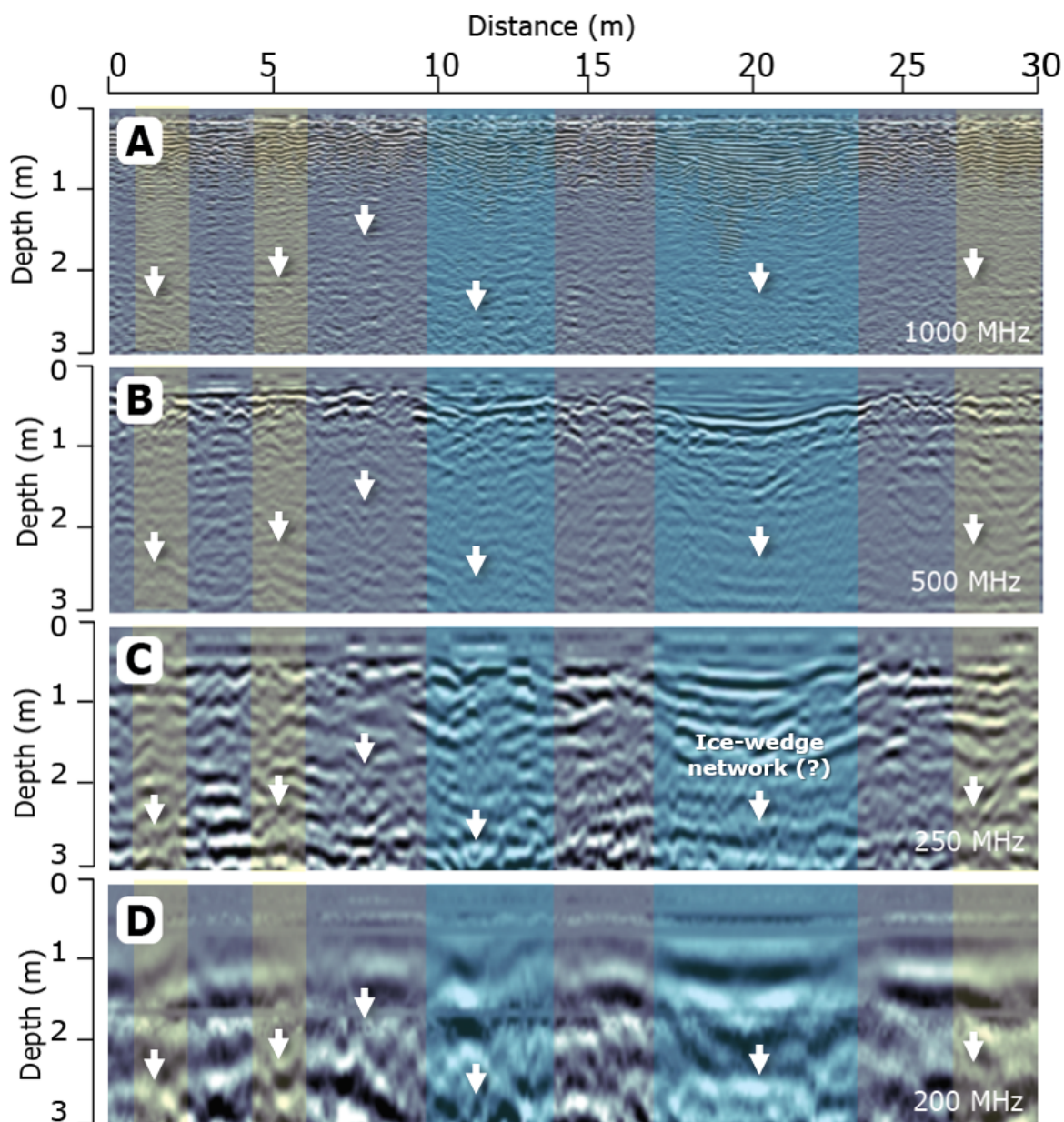


Figure 6.4. Radargram profiles collected across an ice-wedge polygon field at Two Moose Lake (Line Y-7) shown in **Figure 6.5** in survey context, using four GPR frequencies: (A) 1000 MHz, (B) 500 MHz, (C) 250 MHz, and (D) 200 MHz. Two prominent ice wedges are highlighted in blue, with additional potential wedge features marked in yellow. Ice wedges are also marked with white arrows, but most go deeper than 3 m. Higher frequencies (e.g., 1000 MHz) offer finer resolution of near-surface structure, while lower frequencies (e.g., 200 MHz) allow deeper penetration with broader stratigraphic patterns.

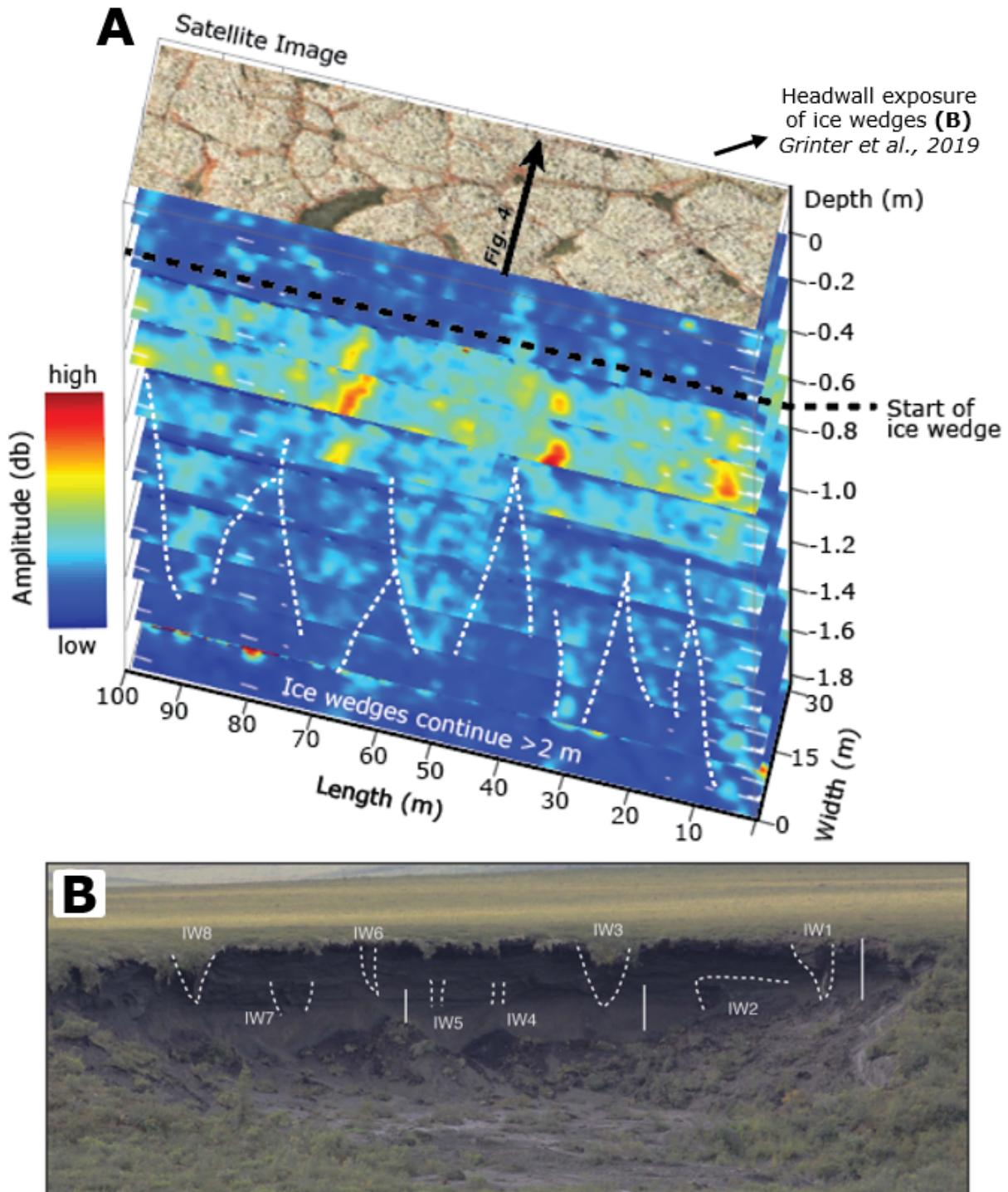


Figure 6.5. (A) Three-dimensional GPR depth slices across an ice-wedge polygon field at Two Moose Lake, with satellite imagery overlain for spatial context. Slices were generated using a velocity of 0.162 m/ns with a thickness of 0.1 m and 50% lateral overlap. While the survey imaged to a depth of 3 m, slices are truncated at ~2 m to emphasize the ice wedge interface beginning at ~0.7 m depth (black dashed line). Dotted white lines outline ice wedge signatures extending below the transition zone below 2m. Arrow indicates orientation of radargrams shown in **Figure 6.4. (B)** A comparison figure from Grinter et al., 2019 showing complex ice wedge stratigraphy exposure in headwall of thaw slump slightly east of the GPR survey draining into Two Moose Lake.

6.3.1.2 SlimSAR

SlimSAR radargrams collected along a 3 km flight line over Two Moose Lake (**Figure 5.6**) reveal subsurface reflectors consistent with stratigraphic boundaries and ice-rich permafrost. The radar system recorded echoes at a sampling rate corresponding to an estimated vertical resolution of ~ 0.3080 m per time sample, and a pulse spacing of ~ 5.07 m derived from GPS data. All four polarizations (HH, HV, VH, VV) show a laterally continuous reflector between 1480-1560 ns (~ 370 -390 sample times), likely marking the base of the active layer or a transition to ice-rich ground. The HH polarization, being co-polarized like GPR, is the most appropriate for comparison and shows the clearest reflection continuity. However, unlike the high-resolution GPR data, we are currently not able to resolve individual ice wedges in the airborne radargrams.

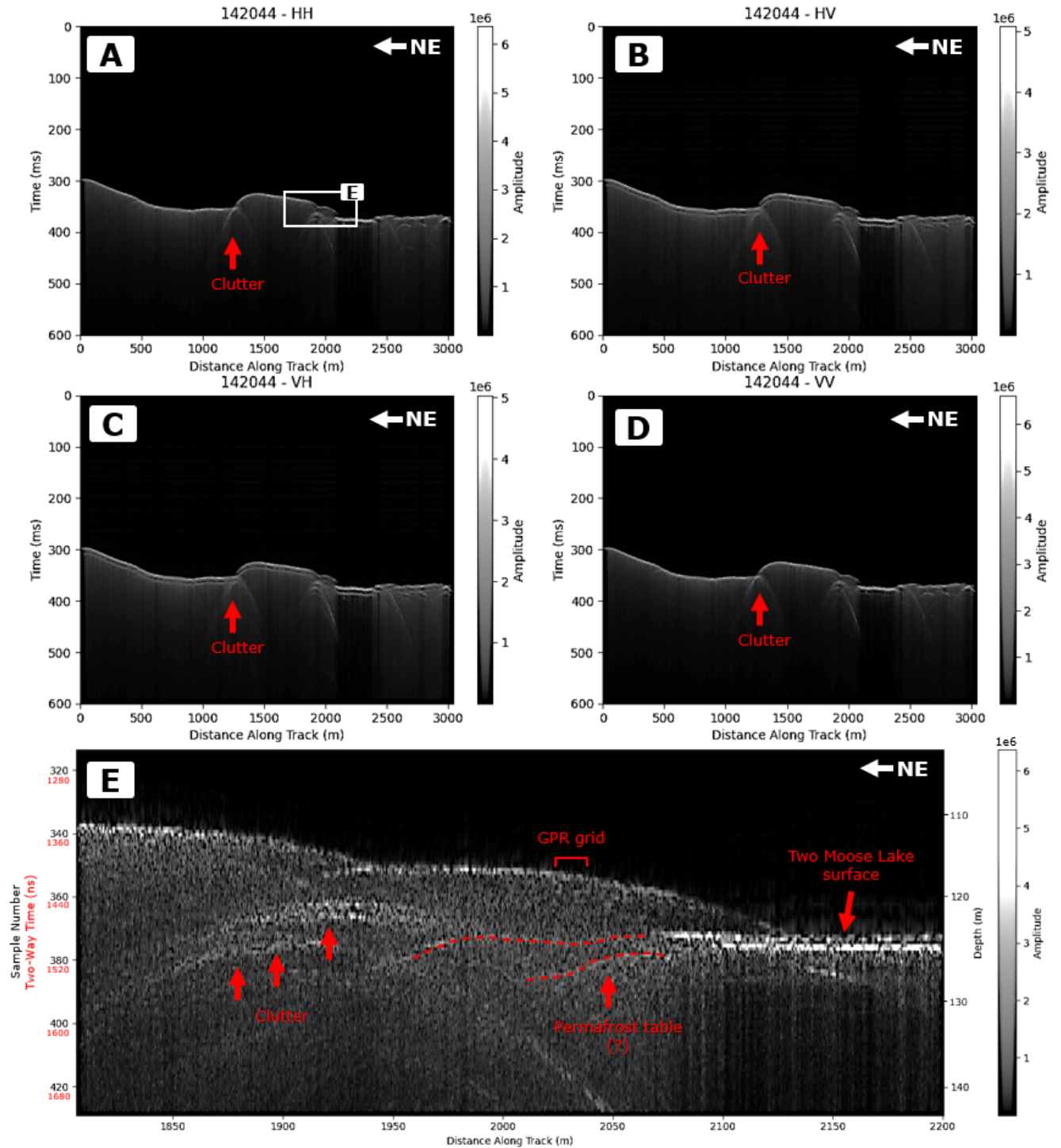


Figure 6.6. SlimSAR radargrams (A) HH, (B) HV, (C) VH, and (D) VV along flight line 142044 over Two Moose Lake, covering ~3000 m of along-track distance and 0–600 sample time. Subsurface parabolas between 300–450 sample time likely represent clutter, but further along the track at ~2000 m may be the base of the permafrost table or bedrock. (E) Zoomed-in section (white box in A) showing GPR grid location and interpreted permafrost table. Differences across polarizations illustrate variations in backscatter and subsurface layering. HH radargram is used as the closest comparison to GPR survey.

6.3.1.3 Sedimentology

The core extracted from an ice-wedge polygon trough at Two Moose Lake (**Figure 6.7**) reveals a five-unit stratigraphy transitioning from frozen, fine-grained soils to massive wedge ice.

Although the uppermost 10 cm of organic-rich active layer was not retained due to coring constraints, frozen ground was present from the start of the frozen ground downward. These units record sedimentary processes, cryogenic modification, and long-term wedge ice formation beneath polygonal terrain:

- **Unit 1 (0.00–0.23 m): Laminated Silty Sand.** This unit represents the shallow frozen active layer and is composed of horizontally laminated silt and fine sand (Sm, Sh, Fm) with rootlets and dark brown colouration. Laminations (2–5 cm) suggest low-energy sheetwash or ponding, common in polygon troughs and/or junctions.
- **Unit 2 (0.23–0.35 m): Silty Clay with Ice Lens.** This interval contains silty clay (Fm) with abundant organic matter and a 4.5 cm dirty ice lens at ~0.25 m (IL), indicating moisture accumulation and refreezing. It represents the lower active layer and upper permafrost transition.
- **Unit 3 (0.35–0.55 m): Deformed Fine Clay.** Grey-green, massive clay (Fm, Fd) with subtle deformation features occupies this interval. Likely formed from freeze-thaw cycling and segregated ice growth within the upper portion of the active layer, this zone reflects dynamic permafrost processes near the freezing front (i.e. free-thaw cycles of >1 year).
- **Unit 4 (0.55–0.73 m): Silty Sand with Organic Matter.** This lighter brown, structureless silty sand (Fm, Sm) is likely reworked material deposited under fluctuating hydrological conditions. It forms the final sedimentary unit above the massive wedge ice.
- **Unit 5 (0.73–1.73 m): Ice-Wedge Ice (IW).** Starting at 0.73 m, this unit consists of massive wedge ice, with alternating clean and “dirty” sections. The presence of dirty ice suggests episodic sediment incorporation during syngenetic (simultaneous) ice-wedge growth. The ice-wedge interface at ~0.73 m corresponds with strong radar reflectors identified in GPR depth slices (**Figure 6.5**), affirming the subsurface continuity of wedge ice. The stratigraphy reveals typical features of polygonal terrain, including progressive soil freezing, vertical ice segregation, and long-term wedge development. These data

provide critical ground-truth for geophysical interpretation and highlight the sedimentological complexity underlying permafrost landforms.

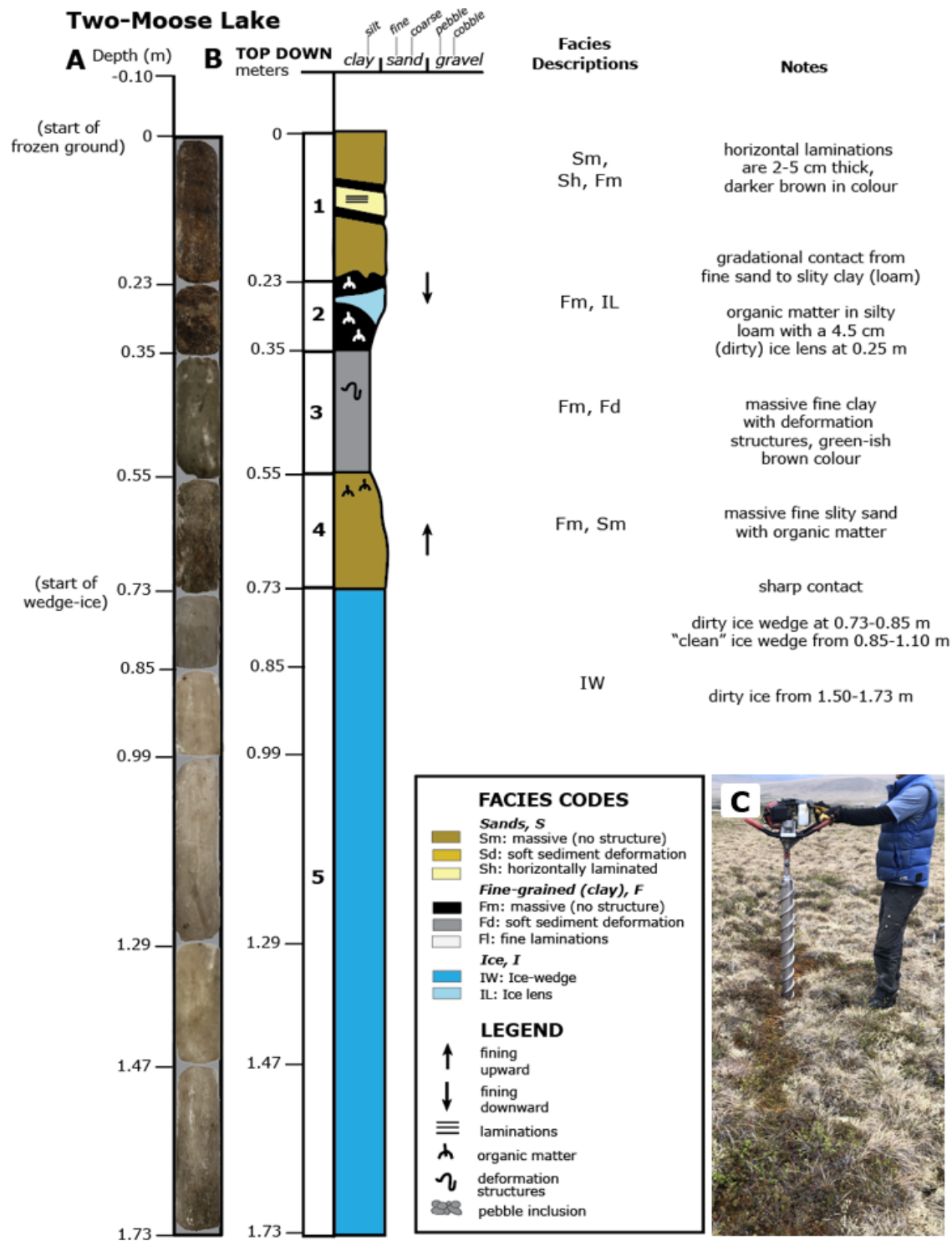


Figure 6.7. Two Moose Lake core profile and sediment log. (A) Photographs of the core (sections missing due to separate packaging and coring constraints). (B) Sediment log with interpreted facies and ice-wedge stratigraphy; frozen ground begins at 0 m. Facies descriptions and notes are outlined in the legend box with facies units 1-5 labelled, (C) Field photo of coring at an ice-wedge polygon trough, where the upper 10 cm remained unfrozen.

6.3.2 Chapman Lake

6.3.2.1 Multi-frequency GPR

GPR profiles collected over the Chapman Lake thaw slump reveal stratigraphic features interpreted as an ice-rich network. The average electromagnetic wave velocity was estimated at 0.154 m/ns ($\epsilon = 3.79$), aligning with values typical of ice-rich permafrost. Despite the potential for deeper penetration at lower frequencies, profiles were cropped to 2 m depth to focus on upper permafrost structure (**Figure 6.8**).

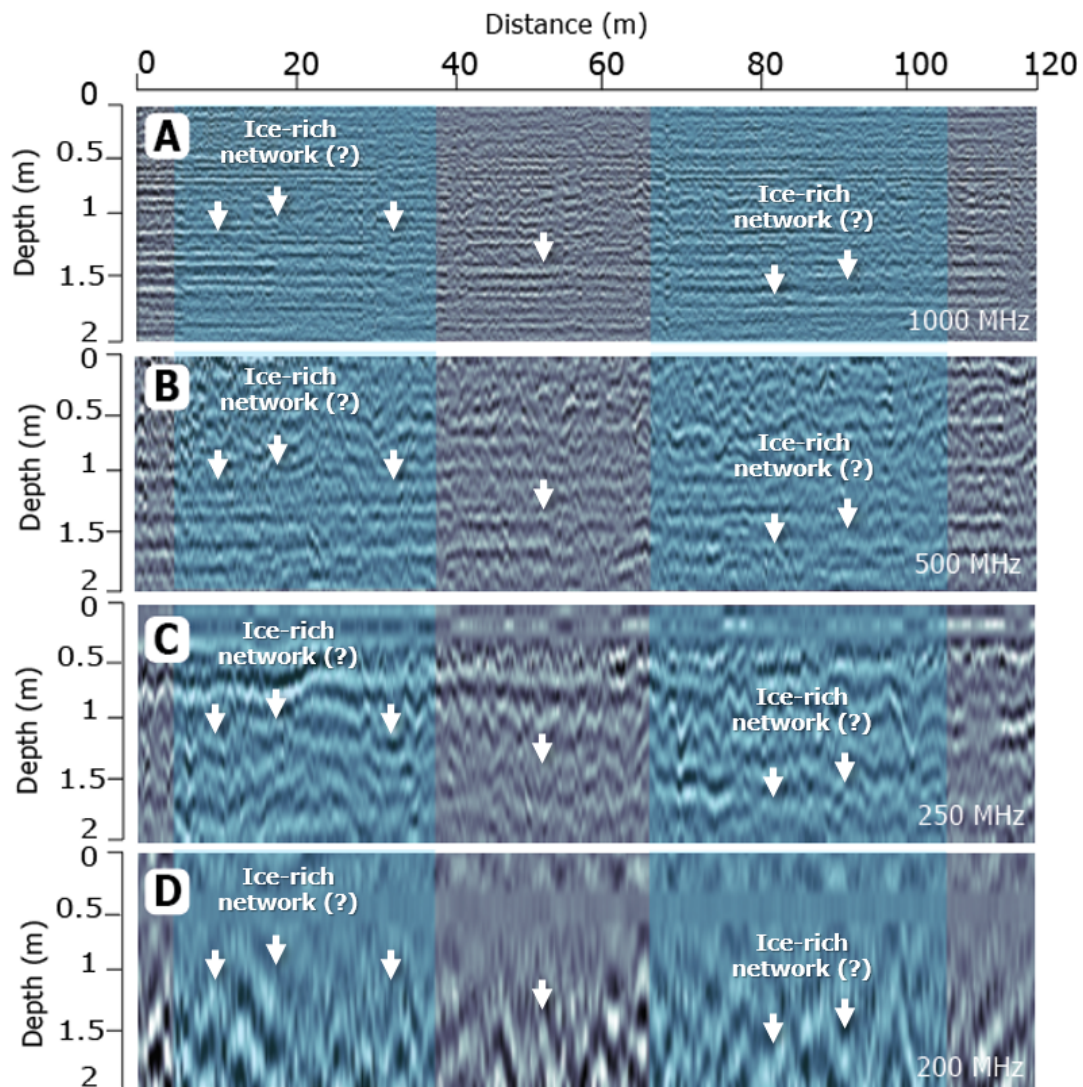


Figure 6.8. Multi-frequency GPR radargrams from Chapman Lake, collected across a re-vegetated thaw slump adjacent to the Dempster Highway. Each panel (A–D) displays GPR reflections from 0–2 m depth for 1000 MHz, 500 MHz, 250 MHz, and 200 MHz frequencies.

The 1000 MHz radargram resolved layering down to 2.0 meters within the active layer. In general, all frequency profiles (especially the 1000, 500 and 250 MHz) captured broad, high-amplitude reflections consistent with sediment over massive ice bodies beginning near ~0.8 m, supporting findings from previous studies that identified this material as relict glacial ice (Dallimore et al., 1992; Sokolov et al., 2020; Yoshikawa et al., 2006). The depths were cropped at 2 m for comparability but the 200 and 250 MHz go until 10 m of penetration depth. These reflectors are consistent with ice exposure documented during initial slump formation in 2007 (Lacelle et al., 2007), though they now lie >0.8 m beneath re-vegetated terrain.

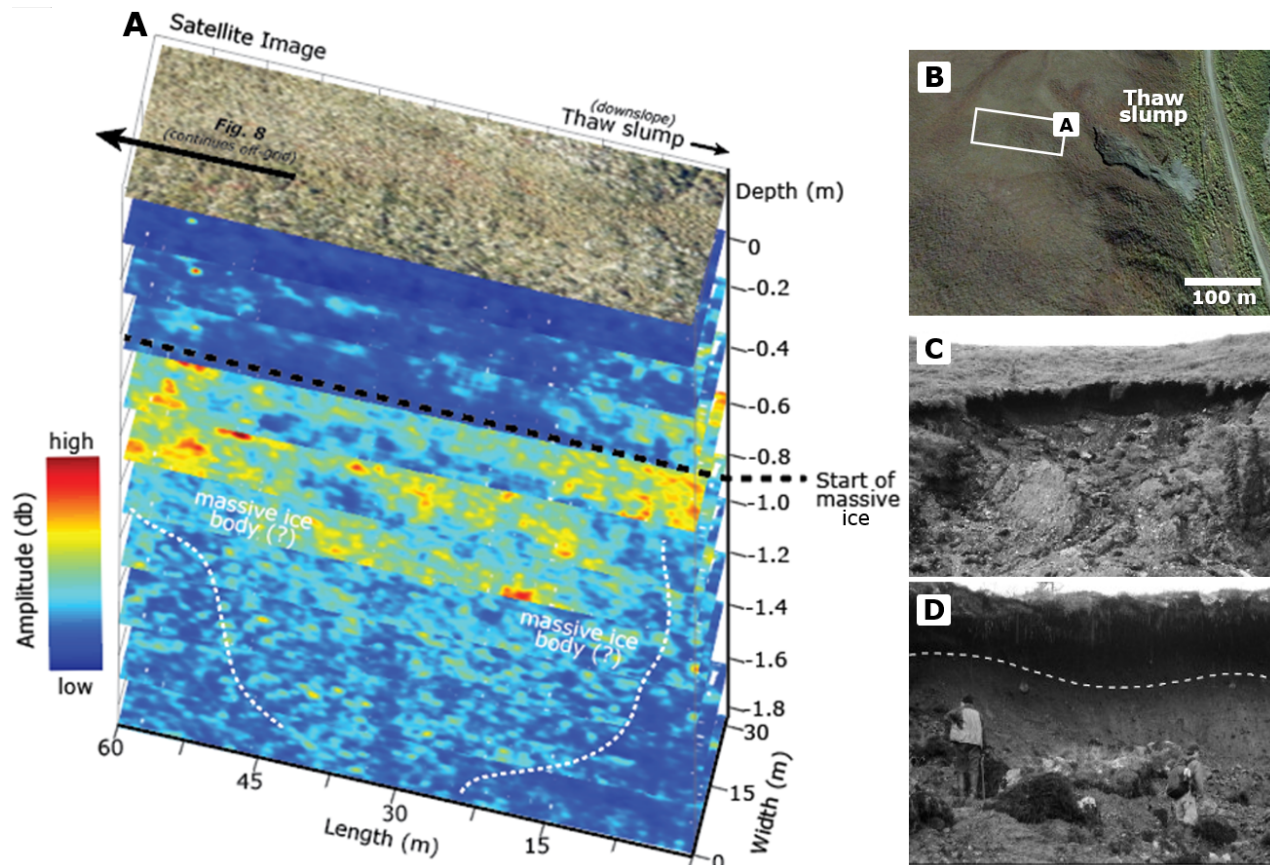


Figure 6.9. (A) Depth-sliced amplitude maps of the Chapman Lake thaw slump derived from 250 MHz GPR data, showing high-reflectivity zones interpreted as massive ground ice. Slices were generated at 0.1 m intervals with 50% overlap and truncated at 1.8 m to highlight ice-rich zones beginning near 0.8 m depth. The mapped interface corresponds to the transition from frozen ground to massive glacial ice. (B) Historical satellite imagery from 2007 of the thaw slump. (C) and (D) are photographs of the massive ground ice exposed in the headwall thaw slump in the Chapman Lake moraine complex on the Blackstone Plateau (taken from Lacelle et al., 2007).

Depth-sliced amplitude maps (**Figure 9**) further support this interpretation. A strong, laterally continuous reflective zone emerges around ~1 m depth, extending across the slump floor and aligning with the mapped headwall boundary. The geometry and reflectivity of these zones are

characteristic of a heterogenous massive ice body (Sokolov et al., 2020) and suggest that the underlying ice body persists beneath the vegetated surface beyond >1 m. Continued monitoring of this site remains important due to its proximity above and adjacent to the Dempster Highway and its role in understanding thaw sensitivity of glacial ice deposits in the region.

6.3.2.2 SlimSAR

SlimSAR radargrams collected along a 3 km flight line over Chapman Lake (**Figure 6.10**) reveal laterally continuous clutter (**Figure 6.10E**) consistent with an artifact from the Dempster Highway road as it was an along-track flight. The system sampled at an interval equivalent to ~0.3080 m vertical resolution, with ~5.03 m pulse spacing based on GPS tracking.

Unfortunately, no distinct ice wedge geometries or ice bodies were resolved. We believe that surface infrastructure or plane geometry may limit interpretation of near-surface radar returns in SlimSAR data.

6.3.2.3 CCR

Capacitively-coupled resistivity (CCR) data collected at the Chapman Lake thaw slump revealed distinct high-resistivity anomalies interpreted as massive ground ice and/or possible ice wedges or highly-saturated pore ice (**Figure 6.11**). Inversion models from all three lines show values exceeding 6000 Ωm , which is consistent with frozen sediments and cryostructures with low clay content (Rossi et al., 2022). Line 1 shows a sharp, localized resistivity peak at ~1.5 m depth, interpreted as an ice wedge, alongside a deeper resistive zone between 3–4.5 m associated with massive ground ice, which we do not see in the SlimSAR data. Line 2 captures a broad resistive body (>7000 Ωm) extending from 1–4 m, indicative of massive, relatively homogeneous ice-rich permafrost. Line 3 highlights multiple shallow ice-rich signals between 1.5–2.5 m and a deeper resistive body >80 m along the transect, consistent with buried massive ice that was investigated at this same site by Grinter et al. (2019). Together, these results align with surface exposures at the slump headwall and support the presence of extensive subsurface ice features.

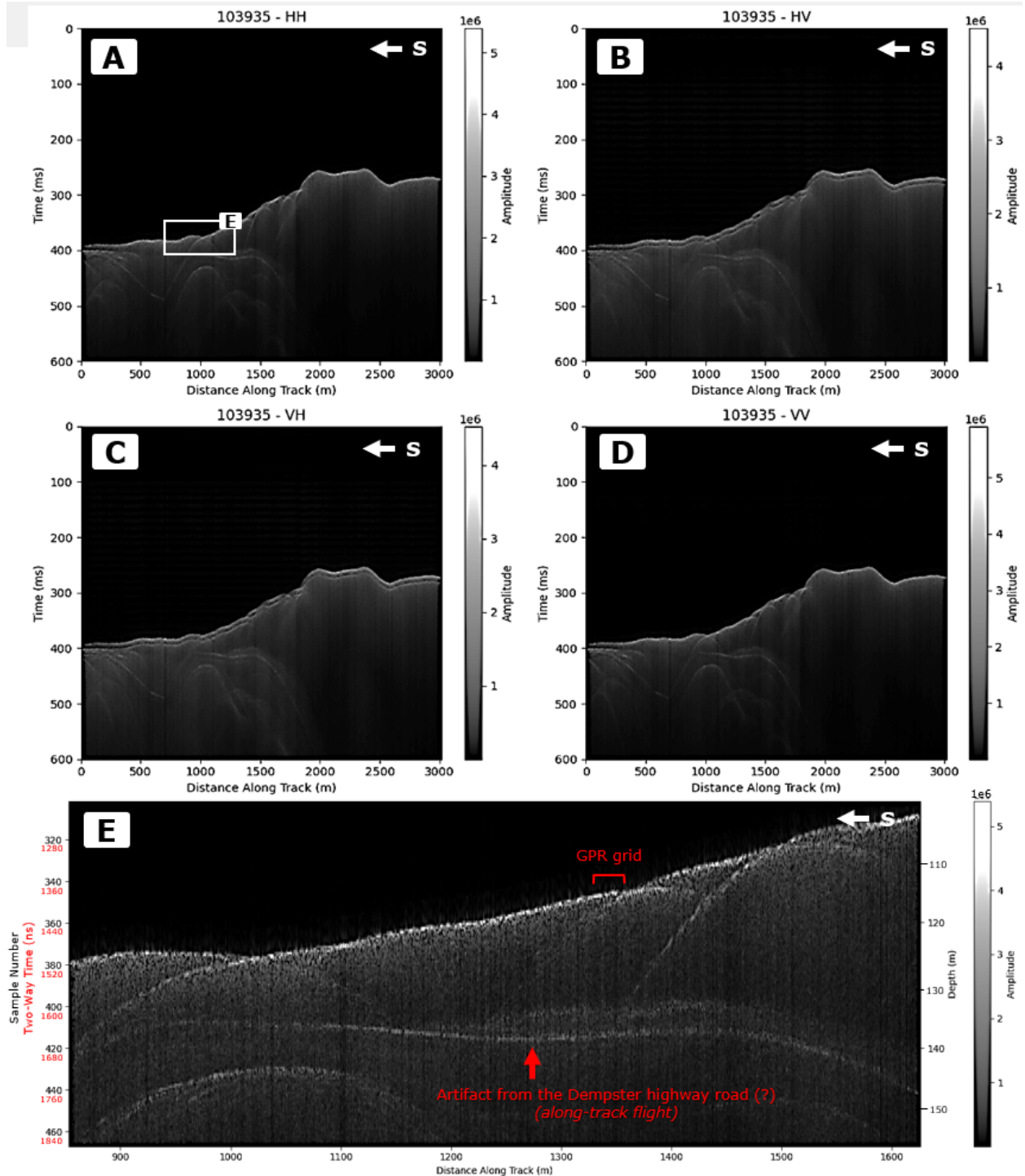


Figure 6.10. SlimSAR radargrams (A) HH, (B) HV, (C) VH, and (D) VV along flight line 103935 over Chapman Lake, spanning ~3000 m of along-track distance and 0–600 ms two-way travel time. Continuous (clutter) reflector is highlighted at 1660 ns (410 sample time). (E) Zoomed-in HH panel highlights the GPR grid and a continuous subsurface reflector, interpreted as the permafrost table. A strong horizontal signal near the surface may represent an artifact caused by the snow-covered Dempster Highway running parallel to the flight line. HH data provide the best structural match to co-polarized GPR.

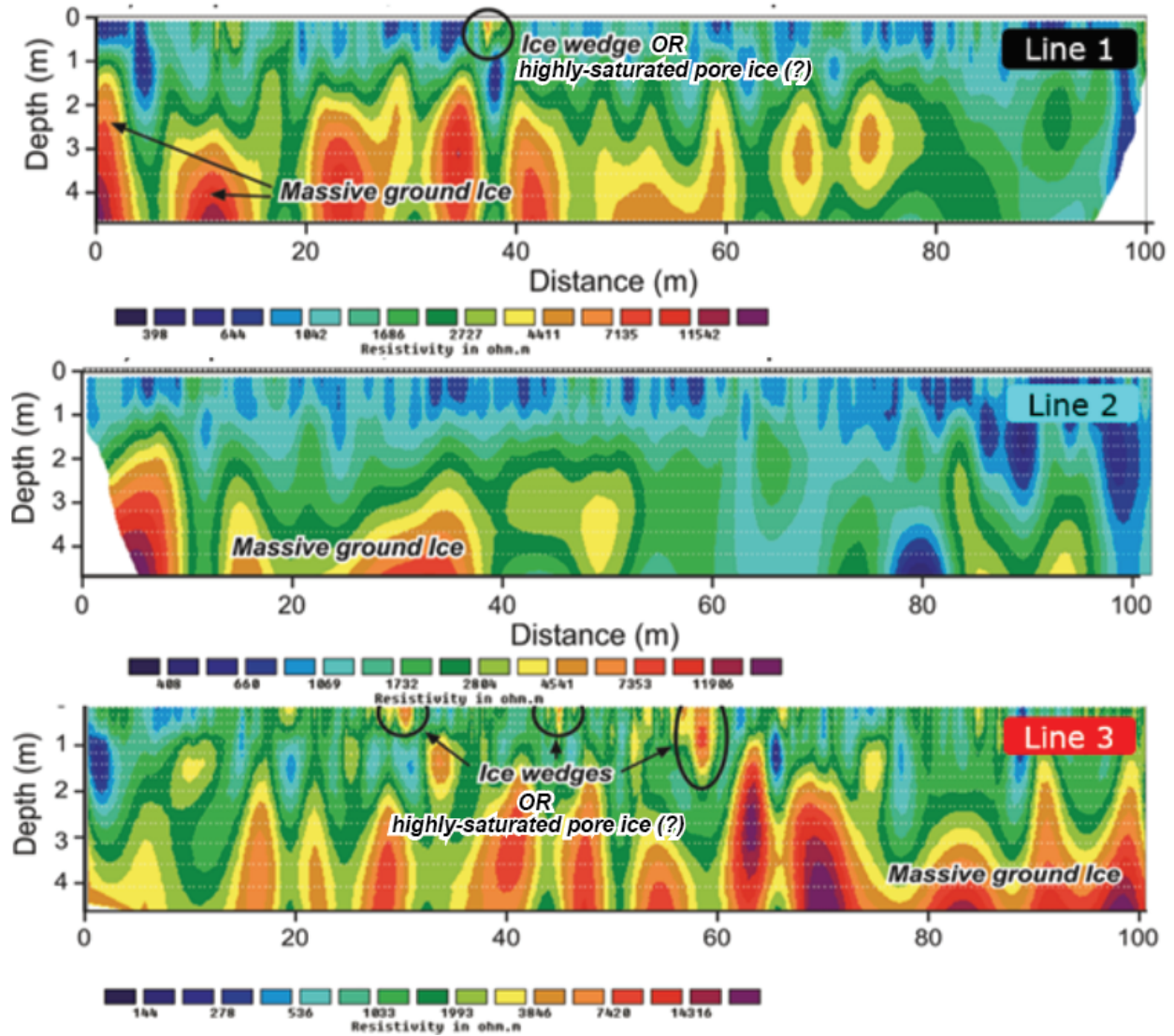


Figure 6.11. Inverted CCR resistivity profiles from the Chapman Lake thaw slump (Lines 1–3), showing high-resistivity zones interpreted as massive ground ice and discrete ice wedges. Resistivity values $>6000 \Omega\text{m}$ are consistent with frozen sediments (after Rossi et al., 2022)

6.3.3 Chapman Lake Airstrip

6.3.3.1 Multi-frequency GPR

Despite being selected as a “control” site due to its flat topography and presumed minimal subsurface complexity, Chapman Lake Airstrip revealed unexpected gravelly and boulder-like signatures. The estimated average EM wave velocity was 0.121 m/ns , consistent with frozen coarse-grained soil ($\epsilon = 6.15$). All radargrams were cropped to 2 m depth for comparability across frequencies although the 500 and 1000 MHz were programmed to “listen” longer up 10m.

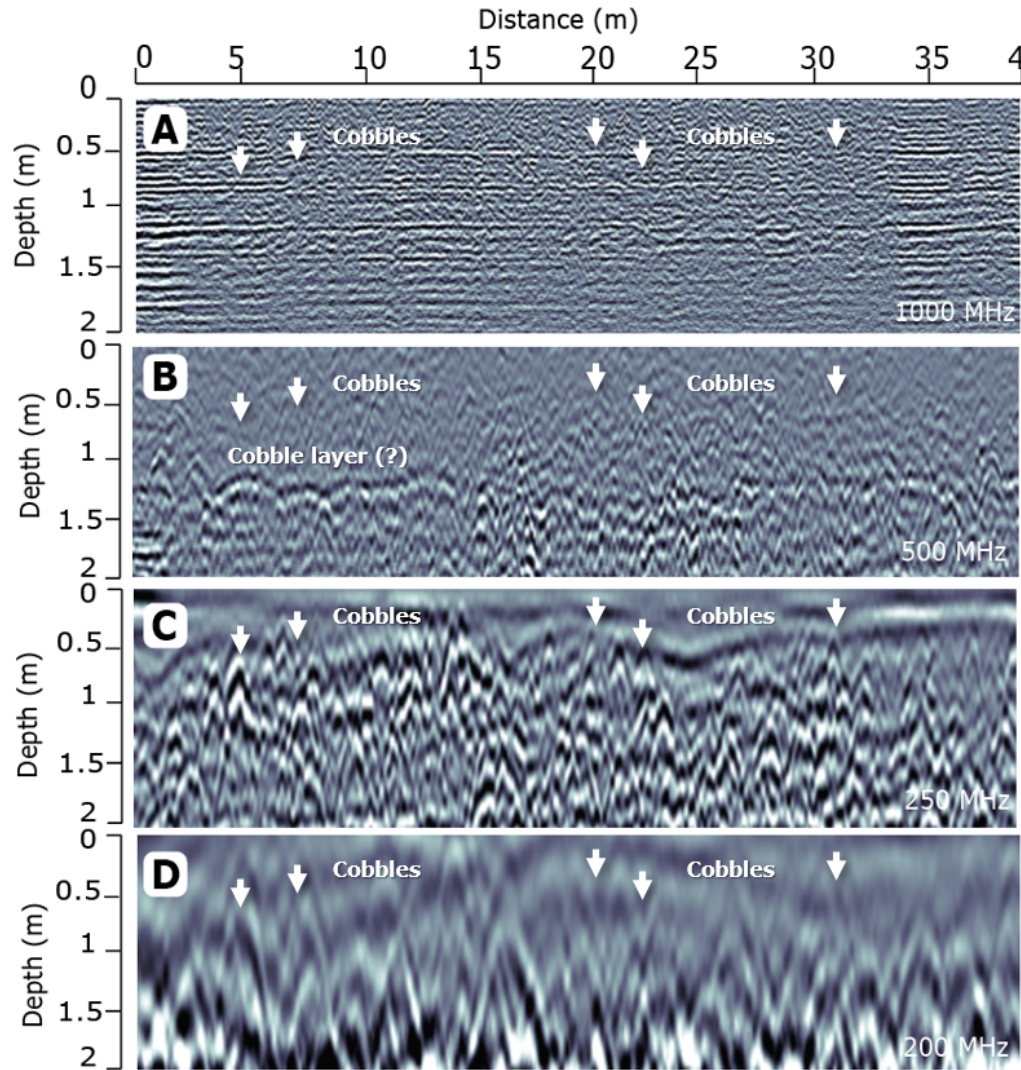


Figure 6.12. Multi-frequency GPR radargrams from Chapman Lake Airstrip showing vertical profiles to 2 m depth. (A) 1000 MHz, (B) 500 MHz, (C) 250 MHz, and (D) 200 MHz profiles. Reflectors are most visible in the 250 and 500 MHz profiles as hyperbolic and dipping returns between ~0.8–1.5 m.

The 1000 MHz profile (**Figure 6.12A**) clearly resolved fine-scale layering down to 2 m. The 500 MHz and 250 MHz profiles (**Figures 6.12B and 6.12C**) offered the best compromise between resolution and depth, showing hyperbolic reflectors and dipping patterns consistent with a gravelly ice mixture between ~0.8–1.5 m. The 200 MHz data (**Figure 6.12D**) exhibited broader, less coherent reflections. Depth-sliced amplitude maps (**Figure 6.13**) revealed high-amplitude anomalies at depths between 0.7 and 1.4 m, trending roughly orthogonal to the road alignment. These are interpreted as possible cobbles and sediment-infill material (likely with pore-ice) from the construction of the airstrip, consistent with the anomalous high-resistivity zones from the YukonGeo report (Camels et al., 2018).

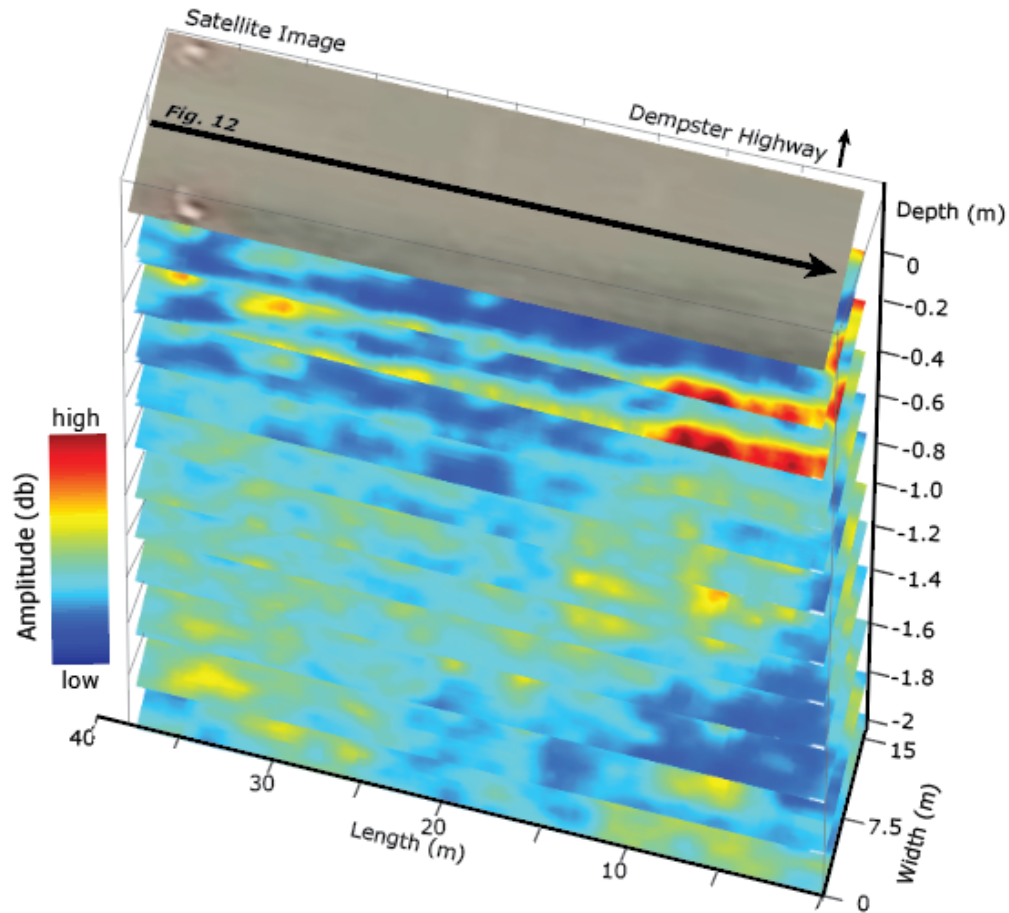


Figure 6.13. Depth-sliced amplitude maps from Chapman Lake Airstrip with 0.1 m thickness and 50% overlap. A satellite image overlays the upper surface for context. Curvilinear, high-amplitude zones between ~ 0.7 – 1.4 m depth align with interpreted gravelly material locations, indicating some permafrost active layer heterogeneity.

6.3.3.2 SlimSAR

SlimSAR radargrams collected along a 3.4 km flight line over the Chapman Lake Airstrip (**Figure 6.14**) reveal a persistent, laterally continuous reflector centered at 840 ns (~ 210 sample time). This boundary is visible in all polarization modes (HH, HV, VH, VV) and is interpreted as the transition from the active layer to underlying frozen ground. The radar system's bandwidth corresponds to a vertical resolution of ~ 0.3080 m, with GPS-derived pulse spacing of ~ 5.68 m. However, gravel stratigraphy and/or individual boulders and boulder-sizes are not clearly distinguishable at this site. This may be due to the coarse vertical resolution of SlimSAR, along-track orientation over the East Blackstone River or Dempster Highway, and the lack of system calibration. Nonetheless, the continuous subsurface reflections suggest a shallow boundary that may be consistent with GPR or could also signify the bottom of either the East Blackstone

riverbed or the sediment-filled subsurface boundary of the airstrip when it was being constructed, both of which are not completely discernable from the SlimSAR radargrams.

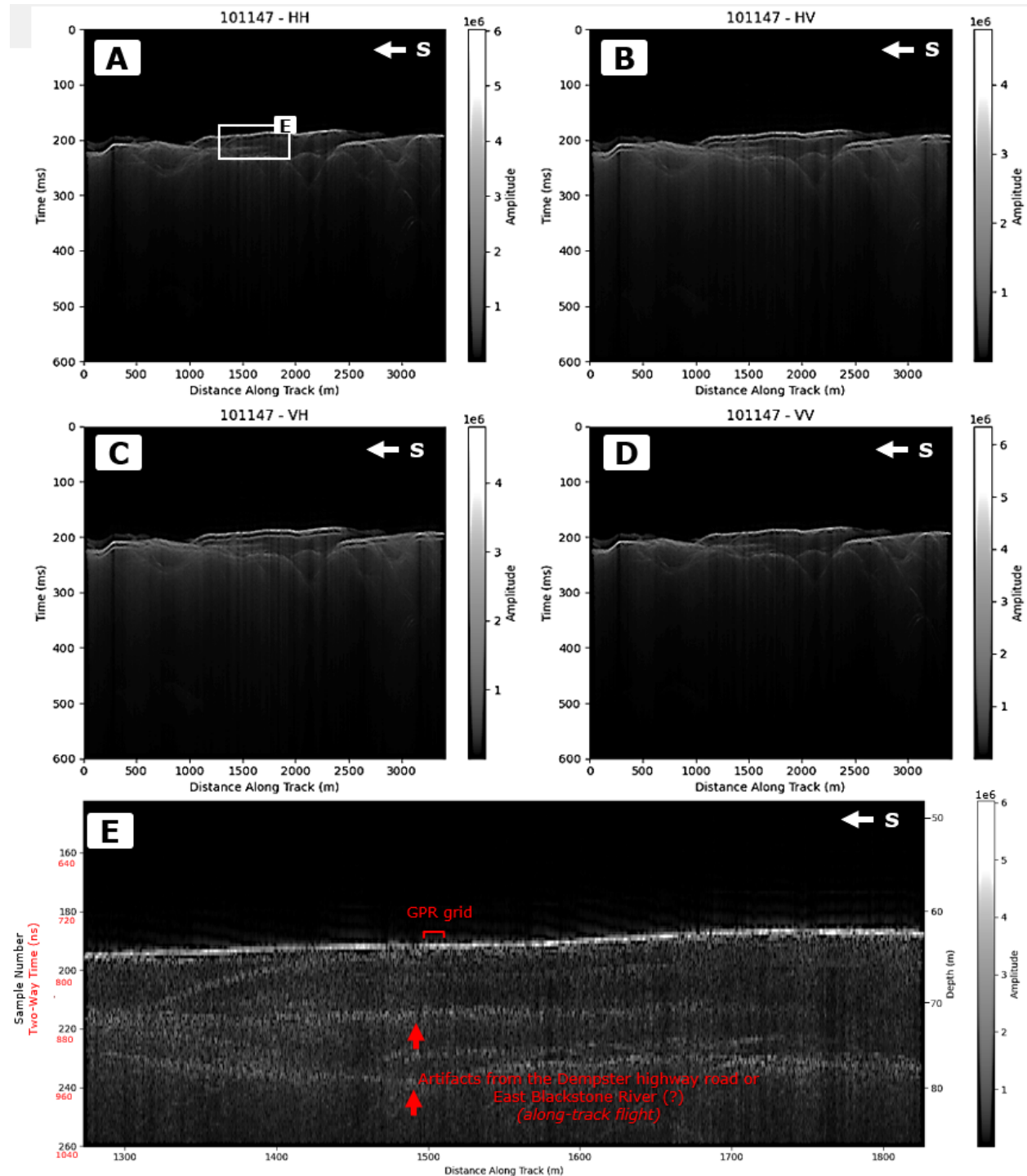


Figure 6.14. SlimSAR radargrams (A) HH, (B) HV, (C) VH, and (D) VV from the Chapman Lake Airstrip site, covering ~3409 m of along-track distance and 0–600 ms two-way travel time. A laterally continuous reflector near 190 and 220 sample time is observed across all polarizations, likely marking the transition from the active layer to ice-rich permafrost. (E) Zoom-in of the area indicated in panel A shows the GPR grid location and strong reflectors

possibly related to the Dempster Highway or East Blackstone River, as the radar sounder was flown along-track. HH polarization provides the clearest comparison to ground-based GPR data.

6.3.3.3 Sedimentology

The 1.41 m core from the Chapman Lake Airstrip site was extracted from the western side of the Dempster Highway, adjacent to polygonal ground, since the airstrip itself is gravel-covered for transportation and inaccessible for coring (**Figure 6.13**). Therefore, this core may not be a direct 1-to-1 ground-truth but could provide context to the subsurface of the airstrip pre-construction regardless. Frozen ground was present from 0.00 m downward. The core reveals three distinct units transitioning from fine-grained sediment to massive wedge ice:

- **Unit 1 (0.00–0.18 m): Silty Sand with Roots.** This uppermost unit consists of fine silty sand (Sm) with visible root material and no internal structure, representing the shallow frozen active layer.
- **Unit 2 (0.18–0.63 m): Clay-Rich Soil with Ice Lenses.** This interval contains silty clay (Sm, IL; Fm, IL) with organic matter and several ice lenses (~0.2 m and ~0.4 m). A hydrogen sulfide odor noted below 0.5 m suggests anaerobic conditions and high organic content.
- **Unit 3 (0.63–1.41 m): Ice-Wedge Ice (IW).** A gradational contact at 0.63 m marks the start of massive wedge ice, with alternating “dirty” and clean zones indicating episodic sediment incorporation during syngenetic growth. On the other side of the highway on the Chapman Lake Airstrip, we could imagine that if given the chance, the ice-wedge would also develop at the same horizon (at ~0.6 m) if it weren’t for the construction of the airstrip itself or the disruption of the ground.

While these stratigraphic results support GPR interpretations of wedge-ice development, a direct one-to-one comparison is limited, as the GPR transect was collected across the snow-covered gravel surface of the airstrip, not the adjacent polygonal ground where the core was obtained.

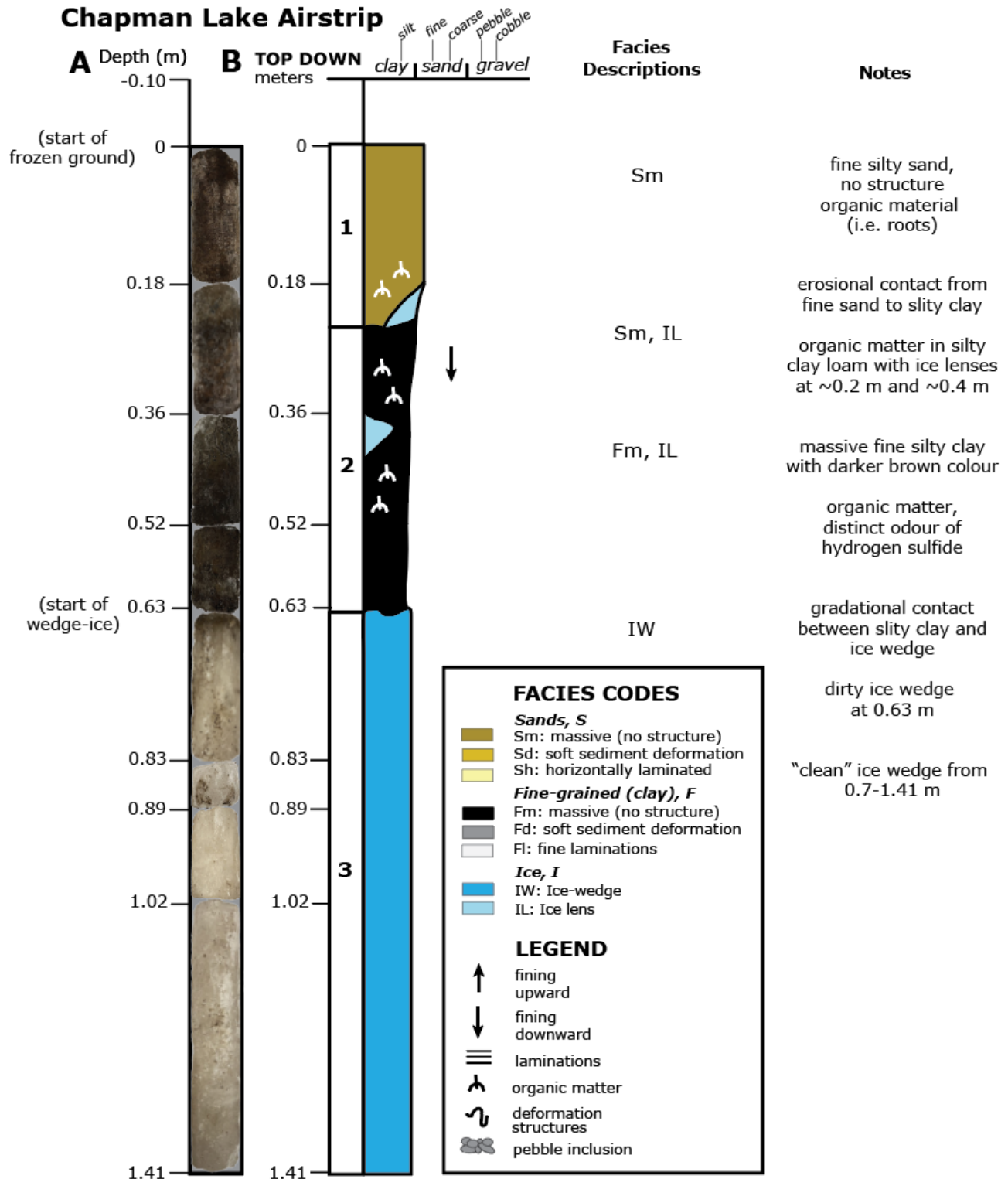


Figure 6.15. Sediment core log and facies interpretation from the Chapman Lake Airstrip area. (A) Core photographs from 0–1.41 m depth, collected from polygonal ground west of the Dempster Highway. (B) Sediment log with interpreted facies, highlighting stratigraphic transitions and the onset of wedge ice at 0.63 m. Ice lenses and organic-rich horizons occur above the wedge ice, with notes indicating distinctive textures and odors. Units 1-3 are also labelled.

6.4 Discussion

6.4.1 Buried Ice Detections from Radar

6.4.1.1 Two Moose Lake: Ice-wedge polygons

At Two Moose Lake, integrated geophysical and sedimentological data confirm the presence of extensive, well-developed ice wedges beneath polygonal terrain. Multi-frequency GPR (250–500 MHz) identified wedge-like reflectors below ~0.7 m depth, with dipping and hyperbolic geometries traceable beyond 2 m, matching curvilinear features in depth-sliced amplitude maps. These GPR signals correspond to the contact observed in a 1.73 m core, which revealed a sharp transition to wedge ice at 0.73 m and alternating bands of clean and sediment-rich ice below, suggesting episodic infill and freezing.

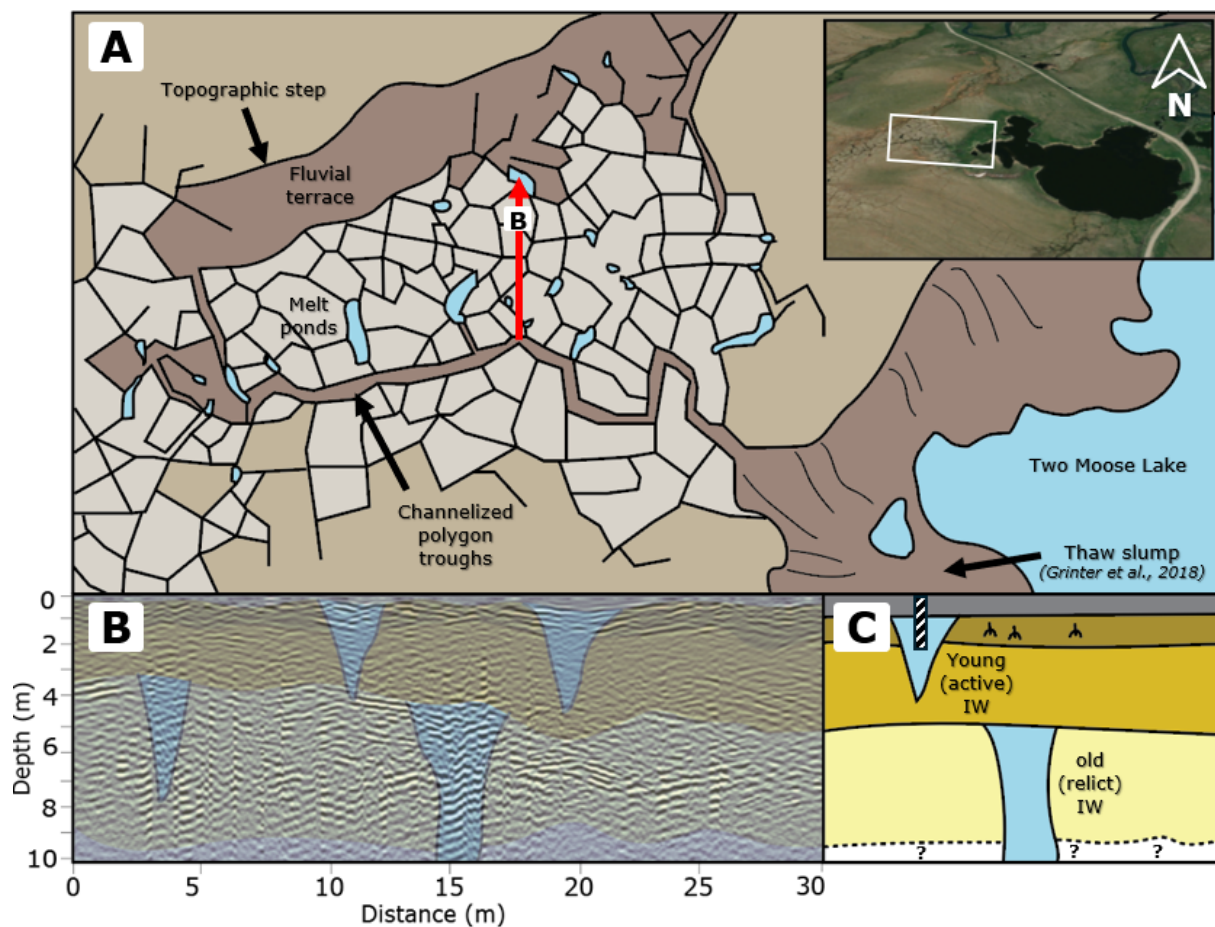


Figure 6.16. Evidence for multiple generations of ice wedges at Two Moose Lake. (A) Map of polygonal terrain with outlined troughs and melt ponds overlying satellite imagery. The red transect line corresponds to the GPR section in panel B. (B) Interpreted GPR profile showing multiple ice wedge structures extending to depths of ~10 m, with overlapping wedge geometries suggesting successive growth phases. (C) Conceptual model of wedge

development at the site, showing young (active) wedges forming above older (relict) wedges within stratified permafrost. Adapted from Grinter et al. (2019).

The alternating clean and dirty ice zones in Unit 5 (**Figure 6.7**) suggest episodic sediment incorporation during syngenetic wedge growth (French & Shur, 2010). The laminated and organic-rich upper units (Unit 1-2) reflect seasonal thaw conditions and accumulation of organic-rich material, while the deformation in Unit 3 may indicate freeze-thaw cycling and ice segregation (Mackay, 1997). The abrupt stratigraphic transitions and variation in wedge geometry seen in the GPR may represent multiple generations of polygon development, consistent with observations by Grinter et al. (2019) on the Blackstone Plateau, where both Late Pleistocene and Holocene wedge activity was recorded at ~2-3 m.

The variation in ice wedge onset depths and geometries, along with differential expression in radar reflectors, may reflect multiple generations of polygon development (**Figure 6.16**). This interpretation is consistent with findings by Grinter et al., 2019, who documented Late Pleistocene and Holocene ice wedge activity on the Blackstone Plateau, including evidence for overprinted or relict polygon networks. These generational signatures, detectable in both radar structure and ice textures, underscore the complex permafrost evolution at this site.

SlimSAR returned a laterally continuous reflector show a laterally continuous reflector between 1480-1560 ns (~370-390 sample times), likely marking the base of the active layer or a transition to ice-rich ground, but lacked the vertical resolution to resolve discrete wedge geometries.

6.4.1.2 Chapman Lake: Thaw Slump & massive ice

The Chapman Lake site is characterized by a retrogressive thaw slump exposing massive ground ice, widely interpreted as relict glacial ice. GPR data revealed high-amplitude, laterally continuous reflectors that suggest the possibility of a highly-saturated ice-rich subsurface as shown by other geophysical signatures of buried massive ice (Dallimore et al., 1992; Sokolov et al., 2020). Unfortunately, SlimSAR radargrams reflectors are mainly interpreted as clutter consistent with an artifact from the Dempster Highway road as it was an along-track flight. The target massive ice body was not resolved, likely due to several factors such as the limited calibration of the airborne system, volume scattering from snow cover, and the flight path's alignment parallel to the Dempster Highway. These GPR observations however, do support

earlier work by Lacelle et al., 2009, who documented the exposure of Pleistocene-age massive ground ice within the Chapman Lake moraine complex. Capacitively-coupled resistivity (CCR) surveys also provided additional support by identifying high-resistivity zones consistent with massive ice. Overall, SlimSAR clutter highlights the challenges of detecting massive ice using uncalibrated airborne radar, especially under snow-covered conditions.

6.4.1.3 Chapman Lake Airstrip: Pore-ice within Gravel Matrix

Although initially selected as a “control” site, GPR and sedimentological data from the Chapman Lake airstrip area revealed no ice-wedge polygons but could possibly be housing frozen coarse-grained sediment made up of construction material and gravels from when the airstrip was being infilled. Although our 1.41 m core extracted west (diagonally across) of the Dempster Highway intersected laminated, ice-rich silt, with wedge ice beginning at ~0.63 m. The ice-wedge may not have had the time to develop on the airstrip due to the change in soil composition due to construction and human interference. However, we do not know how deep the airstrip infilling goes until, but some stratigraphy is reported from observations of borehole DH124 (adjacent to the airstrip) reported by the Yukon Research Centre (Camels et al., 2018), which documented wedge ice from ~0.7 to 3.0 m depth, underlying ice-poor gravel layers. This could mean that there could be wedge-ice beneath the boundary of the bottom of the airstrip.

ERT surveys from that survey report (Camels et al., 2018) also revealed high-resistivity zones in the upper 3–5 m, interpreted as ice wedges, with permafrost thickness ranging between 8 and 15 m. The base of permafrost appears to transition to unfrozen gravel at depth.

Our SlimSAR radargrams over the airstrip show laterally continuous reflectors at 840 ns (~210 sample time; **Figure 6.14E**), but lack the resolution to isolate wedge morphology, possibly due to the adjacent Blackstone River proximity and/or surface heterogeneity.

While we cannot directly align the GPR transect (conducted atop the airstrip) with the core (collected on the west side of the highway), the similarity in wedge initiation depths and stratigraphic structure suggests site-wide presence of ice wedge complexes, but premature development on the airstrip.

6.4.2 Limitations of Buried Ice Detections from Radar

The SlimSAR data utilized in this study were only range-compressed and lacked radiometric calibration due to the absence of dedicated calibration flights. This limitation restricts the quantitative interpretation of backscatter amplitudes, necessitating caution when comparing signal strengths across different polarizations or survey sites. Future calibration efforts are essential to enhance the reliability of amplitude-based analyses.

Both SlimSAR and GPR surveys were conducted in April 2023 under snow-covered conditions, with snow depths ranging from approximately 50 to 80 cm. The presence of snow introduces volume scattering effects, potentially attenuating the radar signals and complicating the interpretation of subsurface features. To mitigate interference, non-metallic snowshoes were employed during data acquisition.

The undulating terrain, especially at Chapman Lake, posed challenges for maintaining consistent GPR antenna orientation and height above ground level. Efforts were made to traverse the terrain as evenly as possible, and GPS corrections were applied to account for elevation changes. However, residual topographic effects may still influence radar data quality and interpretation. Moreover, the GPR system's hardware limitations allowed only four antennas to be daisy-chained simultaneously, necessitating the pairing of antennas (e.g., 500 and 1000 MHz or 200 and 250 MHz). Calibration settings were optimized for the higher frequency in each pair to achieve the best possible resolution. However, this approach may have compromised the depth penetration capabilities of the lower frequency antennas. Additionally, user errors in setting acquisition parameters led to inconsistent depth penetration across some radargrams, highlighting the need for meticulous configuration and validation during data collection.

6.4.3 Implications to the I-MIM Mission

This study directly supports the International Mars Ice Mapper (I-MIM)'s objectives by testing the detectability of shallow ground ice using both airborne and ground-based radar systems in analogue permafrost environments. While SlimSAR L-band radar provided laterally continuous stratigraphic reflectors (with some mostly clutter), it did not resolve discrete ice wedge structures, likely due to limited resolution, lack of calibration, and snow-induced scattering. This

outcome emphasizes the need for refined airborne radar design and calibration strategies to meet I-MIM's requirement of detecting decameter-scale ice bodies within the upper $\sim 1\text{--}10$ m.

In contrast, ground-based GPR at 250 and 500 MHz consistently resolved ice wedge morphology and stratigraphic transitions within the upper 2–3 m, highlighting the critical role of higher-frequency radar systems in near-surface ice mapping. Our results suggest that a dual-frequency radar strategy may be optimal for Martian applications, balancing penetration depth with the resolution needed to detect complex cryostructures. CCR added complementary constraints on subsurface ice distribution, highlighting the value of integrating dielectric and electrical property measurements. Collectively, these findings reinforce I-MIM's goal to improve subsurface ice detection methodologies and advocate for pairing orbital radar with surface-based or deployable systems and geophysical validation tools. Future work, including cross-polarization analysis of SlimSAR and expanded calibration datasets (Mishev et al., in prep), will help refine radar backscatter models and guide the selection of polarization modes and operational frequencies in support of I-MIM science objectives.

6.5 Conclusion

This study presents a multi-instrument geophysical investigation of permafrost and ground ice conditions across three sites in Tombstone Territorial Park, Yukon. We acquired ~ 9 km of airborne SlimSAR radar data along three 3-km flight lines, ~ 180 individual GPR transects across the three grids, two shallow (≤ 1.7 m) sediment cores, and one 3-line CCR survey at the Chapman Lake thaw slump. GPR surveys were conducted using the common-offset method over grids measuring 60×100 m at Two Moose Lake ice-wedge polygons, 30×60 m at Chapman Lake thaw slump, and 15×40 m at the Chapman Lake Airstrip. These results were supplemented with existing ERT data from Site DH124 at the airstrip (Camels et al., 2018) and CCR data for the Chapman Lake thaw slump. This integrative approach allowed for site-specific comparison of radar, resistivity, and sedimentology methods and provided high-confidence cryostratigraphic interpretation across a range of ice-rich landforms.

GPR at 250 and 500 MHz consistently resolved subsurface ice wedge morphology and cryostratigraphic boundaries to depths of ~ 2.5 m, with radar velocities ranging from 0.162, 0.154, and 0.121 m/ns. At Two Moose Lake, ice wedge contacts identified in depth slices

(~0.7 m) aligned precisely with sediment core observations and corresponded with late Holocene wedge development (Grinter et al., 2019). At Chapman Lake, radargrams and CCR inversions identified broad, high-amplitude reflections and high-resistivity anomalies ($>7000 \Omega\text{m}$) consistent with relict massive ice, supporting earlier interpretations by Lacelle et al. (2009). GPR and coring at the Chapman Lake Airstrip revealed unrecognized frozen gravel complexes and an ice-wedge beginning at ~0.63 m (across the airstrip), confirming the shallow permafrost heterogeneity inferred in the DH124 borehole and ERT data (Camels et al., 2018). Ice-wedges may not have been able to develop on the airstrip due to human interference of the airstrip construction back in the 1970s.

SlimSAR radargrams collected under snow-covered conditions showed some reflectors, interpreted as the permafrost table or transition to ice-rich sediments with medium-confidence. However, the system lacked sufficient resolution to resolve the contrast between discrete pure-ice wedge versus highly saturated pore ice geometries as the GPR was able to do. Signal attenuation from snow, absence of radiometric calibration, and along-track orientation parallel to infrastructure (e.g., Dempster Highway) further limited interpretation. While SlimSAR was valuable for delineating potential dielectric boundaries, further airborne validation and instrument calibration was essential for identifying and characterizing fine-scale cryostructures.

Beyond supporting the objectives of the International Mars Ice Mapper (I-MIM) mission, these results contribute directly to terrestrial applications, particularly for infrastructure risk assessment along the Dempster Highway. The presence of subsurface permafrost and buried glacial ice (within 1–2 m of the surface) within the three sites underscores the vulnerability of the region to thaw-related deformation. These findings emphasize the need for continued ground-truth geophysical monitoring and suggest that integrated radar and resistivity methods can meaningfully improve hazard prediction in permafrost corridors. For Mars analogue studies, this work demonstrates the potential of combining L-band radar and high-frequency GPR to optimize subsurface ice detection under varying terrain, stratigraphic boundaries, and thermal conditions.

CHAPTER 7

7 Synthesis & Future Work

This chapter synthesizes the scientific findings presented in Chapters 4, 5, and 6, connecting them to each other and to the broader literature. Section 7.1 focuses on radar-based discoveries from SHARAD in Phlegra Montes (Chapter 4). Section 7.2 integrates the geomorphological analysis of Phlegra Montes (Chapter 5), while Section 7.3 connects insights from the Yukon fieldwork to the proposed International Mars Ice Mapper (I-MIM) mission (Chapter 6). Finally, Section 7.4 outlines remaining open questions, future research directions, and new hypotheses that emerged during this work but fall outside the scope of the current thesis.

7.1 Radar Detections of Debris-covered Glaciers in Phlegra Montes, Mars (Chapter 4)

7.1.1 Novel SHARAD Equation

In this thesis, I developed a novel analytical application to investigating *two-layered dielectric models* that explicitly includes debris-lag thickness overlying Martian glaciers using the SHARAD radar equation (Equation 2). Previously, Petersen et al. (2018) adopted a uniform bulk dielectric constant ($\epsilon \approx 3.2$), assuming an ice-rich composition. However, these assumptions ignore the spatial distribution of lithic surface lag material and assume a constant (mean) lag thickness throughout the surface of the VFF.

Building on Gallagher et al. (2021)'s preliminary radar work on Phlegra Montes in combination with other research such as Holt et al. (2010), Nunes et al. (2011), and Smith and Holt (2010) who began to address the significance of depth correction, I expanded this approach by explicitly solving for the debris thickness in a debris-ice composite system (**Figure 4.2**). This was done by calculating bulk dielectric constants (assuming dielectric values for both ice and regolith), then calculated lag thickness (Equation 8). This refinement revealed that ϵ' values >3.2 provide better geometric fits when comparing radar-derived depths to local topography (**Figures 4.4 - 4.11**). These improvements suggest that Phlegra Montes hosts thicker debris covers than previously thought, potentially exceeding the $\sim 1\text{-}10$ m SHARAD resolution in some areas, and imply a

combination of both varying sublimation stages due to differential lag and/or more debris delivery.

This analytical model aligns with other studies showing that lag buildup can thermally insulate underlying ice and affect radar wave propagation (Bramson, 2018). It also enhances the reliability of SHARAD for mapping debris-covered glaciers or VFFs by accounting for variable surface cover in dielectric models. By analytically estimating the radar delay caused by a (~1-10 m) dry debris lag, this approach improves depth estimations and helps distinguish between thick ice bodies and thinner, lag-covered deposits. As a result, it strengthens the case for using an analytical method for lag thickness estimation with SHARAD across Mars' mid-latitudes.

7.1.2 Discovery of a Terraced Glacier

SHARAD detections in Phlegra Montes identified a glacier system with two sub-parallel basal reflectors (Site 2; **Figure 4.5**), marking the first documented observation of such terraced basal glacial topography on Mars. The reflectors exhibit consistent geometry across multiple radar lines and were validated using clutter simulation mitigation and depth correction, reducing the likelihood of surface clutter artifacts (**Figure 4.5D**). This finding builds on earlier radar studies (Holt et al., 2008; Petersen et al., 2018c; Plaut et al., 2009), who documented basal reflectors in Martian VFFs, but not internal terracing. The presence of terraced internal structure expands current understanding of Martian VFFs, suggesting that at least some glacial systems in Phlegra Montes may have undergone more complex thermal and mechanical evolution than previously recognized, including possible transitions between cold- and warm-based states (i.e. Gallagher et al., 2021).

Current SHARAD resolution and 2D geometry limit our ability to fully map all VFFs in Phlegra Montes along with other physical limitations (i.e. complex topography; more details in Section 4.2). Future work using 3D SHARAD imaging and/or higher-resolution DEMs (i.e. HRSC or CTX DTMs) for clutter simulation as well as high-frequency radar systems (i.e. from a landed platform or orbiter) would greatly improve our ability to trace reflector continuity and cross-sectional relationships in rough terrain. In particular, a focused 3D radar mapping campaign of Phlegra Montes, similar to the techniques done by Perry et al. (2024) for Deuteronilus Mensae, could constrain reflector geometries and help differentiate stepped basal topography from

perhaps the possibility of an internal ice-sediment interface, which is unlikely. Understanding whether this VFF has other complex subsurface basal topography and reflectors is a key ingredient to interpreting glacier formation and processes on Phlegra Montes.

7.2 Glacial landsystem analysis and geomorphology of Phlegra Montes, Mars (Chapter 5)

7.2.1 Methods in Mapping & Naming Planetary Landforms

To address the heterogeneity in viscous flow features (VFFs) and avoid ambiguous nomenclature (e.g., LDAs, LVFs, CCFs or incised CCFs that become LDAs), this study developed the Confinement Index (CI) as a quantitative metric for characterizing glacier confinement across Phlegra Montes. Rather than classifying landforms based solely on visual morphology, the CI measures the percentage of a glacier's boundary in contact with rising topography, enabling a continuous spectrum of classification from unconfined ($CI = 0$) to fully confined ($CI = 1$).

This approach builds upon prior geomorphological mapping efforts (e.g., Levy et al., 2014; Milliken et al., 2003) but introduces a reproducible, GIS-based method to quantify glacier-topography interactions. Using high-resolution CTX imagery overlaid on MOLA/HRSC digital elevation models (DEMs), VFF polygons were segmented into confined and unconfined boundaries. Polygons were categorized using slope directionality, and zonal statistics extracted confinement metrics for each mapped VFF.

By focusing on confinement rather than inherited (or potentially ambiguous) terminology, this method enables a more objective, quantitative, and continuous comparison between Martian and terrestrial glaciers, supporting finer-scale geomorphological analysis. It also offers a reproducible framework for automated classification in future missions or large-scale remote sensing surveys. However, this approach is more time-intensive than traditional visual classification schemes (i.e. labeling features as CCFs or LDAs), making it less practical for global-scale mapping or rapid mapping tasks. The CI method is most valuable in regional studies aimed at quantifying landform-process relationships, testing specific hypotheses about glacier dynamics, or enabling AI and machine learning classification workflows. It does not require topographic data, relying instead on planform geometry from imagery (i.e. CTX or HiRISE), which makes it broadly applicable even in areas with limited elevation coverage. Thus, CI

should be viewed as a complementary tool, best deployed when classification consistency and reduced subjectivity are priorities.

7.2.2 Relationship and Latitudinal Trends of the LDM and VFFs

The relationship between the Latitude Dependent Mantle (LDM) and VFFs in Phlegra Montes reveals strong latitudinal trends and stratigraphic transitions that reflect differing formation epochs and depositional environments. Historically, the LDM and VFFs were considered separate phenomena (Lucchitta, 1984; Squyres, 1979, 1978), but recent work has questioned whether they are part of an interconnected cryospheric system (Baker and Carter, 2019; Hepburn et al., 2020b). This study supports the interpretation that the LDM is stratigraphically and temporally distinct from the VFFs, based on differences in surface texture, topography, and radar geometry. VFF-associated textures (Knobby and Lineated terrain atop the VFF) are consistently buried by younger LDM-associated terrains (Smooth-Undulating and Scalloped atop the VFF and plains), particularly at higher latitudes. For example, **Figure 4.13** demonstrates how scalloped terrain often follows the buried patterns of Lineated terrain, implying overprinting rather than co-evolution.

Moreover, morphological transitions along latitude suggest a progression from a relict glacial landscape at latitudes $<40^{\circ}\text{N}$, to a mantle-dominated regime north of 48°N . Massif-apron ratios, centroid offsets, and SHARAD-derived ice thicknesses corroborate this trend, with increased burial and ice preservation at higher latitudes. These results align with climate models indicating poleward migration of the stable near-surface ice line under current low obliquity (Laskar et al., 2004; Madeleine et al., 2009). Altogether, these findings support a landsystem transition: VFFs formed during earlier, high-obliquity epochs that were buried by subsequent LDM emplacement. This marks a shift from dominant glacial processes of VFFs (i.e. snow accumulation, active glacial flow under debris, and ice deformation) to more recent emplacement of the LDM reflecting a shift to possibly colder, drier conditions with limited precipitation and ongoing sublimation-driven modification.

Phlegra Montes, a mountain range that covers more than 20° of latitude in the icy mid-latitudes of Mars, thus offers a compelling case study for testing models of Amazonian ice stratigraphy and climate evolution. Future work could combine radar, spectral, and high-resolution

topographic data to resolve mantle thicknesses and their variation with latitude, test burial ages via crater counting (more on Sections 7.4.1 and 7.4.2), and constrain volatile redistribution pathways during obliquity-driven climate cycles. Ultimately, this work informs both models of mid-latitude ice stability and the ongoing debate over the origin and evolution of Mars' Amazonian cryosphere.

7.3 The Case and Significance of a Future Radar and Ice-focused Mission on Mars (Chapter 6)

The results of this thesis, and the tantalizing interpretations that cannot be fully confirmed with current instrumentation, demonstrate that future radar and ice-focused missions on Mars are critical for advancing our understanding of near-surface cryospheric processes, ice stability, and water resource potential. This conclusion is informed not only by SHARAD-based discoveries in Phlegra Montes (Chapter 4), but also by the Earth-based geophysical field investigations in Tombstone Territorial Park, Yukon (Chapter 6), which served as a direct analogue for Martian ice-rich terrains.

In the Yukon Territory, ground-based GPR, SlimSAR airborne radar, and resistivity surveys validated the importance of radar frequency, resolution, and platform scale for detecting shallow ground ice (1–5 m depth). For instance, high-frequency GPR (250–500 MHz) was optimal for resolving ice wedges and shallow stratigraphy (**Figure 6.4**), while the L-band SlimSAR in sounding mode was not able to resolve shallow (1–10 m) subsurface ice targets possibly due to plane geometry, topographic artifacts, and the need for further data calibration. These findings are a work in progress and may offer a proof-of-concept for Mars radar design, demonstrating the value of multi-frequency, multi-scale radar systems in ice detection. Together, these terrestrial and planetary datasets support the case for a mission such as the International Mars Ice Mapper (I-MIM), which would provide orbital, high-resolution radar imaging of mid-latitude ice deposits. A mission of this kind would fill critical knowledge gaps by: (1) mapping shallow ice thickness (1–10 m) and layering at sub-1 m vertical resolution; (2) identifying accessible water-ice for potential in-situ resource utilization (ISRU); (3) constraining climate and glacial history through detecting internal stratigraphy and 3D ice distribution, and (4) supporting site selection for human missions and robotic landers

This thesis thus supports the scientific and strategic rationale for a radar-focused mission, bridging planetary science and Earth-based analogue validation. Phlegra Montes stands as a key candidate region for such a mission, given its accessible latitudes, high ice potential, and well-preserved glacial landforms.

7.4 Insights into Future Directions and Open Questions

Building on the findings presented in this thesis, several future research directions have emerged that aim to deepen our understanding of Martian glacial systems and their Earth analogues as well as technologies used in tandem with geophysical and geomorphological investigations. These future questions and projects combine expanded remote sensing analysis, process modeling, and field-based validation using terrestrial analogues.

7.4.1 ISSM or IGM on Phlegra Montes

To better understand the flow behavior and accumulation history of Martian glaciers, especially in a topographically complex regions like Phlegra Montes, future work could also involve implementing an Ice Sheet and Sea Level System Model (ISSM) or custom Instructed Glacier Models (IGM) on Phlegra Montes. ISSM is a sophisticated, open-source, finite-element model developed by NASA/JPL and UC Irvine for simulating the thermomechanical behavior of ice sheets and glaciers. It is widely used to model ice flow, basal processes, and climate-ice interactions on Earth (Cuzzone et al., 2019; Larour et al., 2012; Schlegel et al., 2018, 2013) and increasingly in Mars ice studies for forward modeling of VFFs and polar ice (Arnold et al., 2022; Butcher et al., 2023; Fard et al., 2024; Fard et al., in review; Smith et al., 2022). Similarly, IGM is a 3D ice flow dynamics model that uses a Convolutional Neural Network (CNN) informed by physics-based constraints to simulate glacier evolution (Leclercq et al., 2023). Both ISSM/IGM are plausible models to use, but more research could be done to see which would be more appropriate, especially for understanding how debris thickness can affect flow rates (Anderson and Anderson, 2025).

While the absolute accuracy of such models may be limited by uncertainties in SHARAD-derived ice thicknesses, lag properties, and past Martian climate inputs, they remain powerful tools for hypothesis testing. For example, by using realistic inputs for obliquity-driven accumulation and ablation rates over a glacial cycle, we could explore which combinations of

input parameters result in present-day ice geometries consistent with radar and geomorphology observations in Phlegra Montes from Chapters 4 and 5. Even if multiple parameter sets can yield similar outputs, such modeling can help constrain plausible formation scenarios (for both VFFs and/or LDM) and timescales for these landforms. These models could also simulate basal stress regimes, flow velocities, the role of debris insulation, and long-term ice evolution under varying obliquity scenarios, all of which would be otherwise unresolvable with current remote sensing. Ultimately, using ISSM or IGM on Phlegra Montes would help identify key physical processes driving glacier preservation on Mars and prioritize data needs for future missions aimed at resolving ice geometry, age, and (potential) internal stratigraphy.

7.4.2 Multi-generational VFFs in Phlegra Montes

In Chapter 5, I explore the geomorphological and latitudinal trends of VFFs in Phlegra Montes. While we focused on “simple” VFF systems (i.e. isolated LDAs that were not superposing other VFFs), future work should focus on constructing a chronostratigraphic framework for Phlegra Montes by integrating crater counting to date surface ages with geomorphic mapping of overlapping VFF units. Ideally, the crater size-frequency distribution (CSFD) analysis will estimate surface age based on the number and size distribution of impact craters, under the assumption that older surfaces accumulate more craters over time (Michael and Neukum, 2010).

A key goal would be to determine the timing, frequency, and climatic context of past glacial events outlined in Section 3.2.2. While CSFD on VFF surfaces can help estimate the age of the most recent resurfacing event, it does not capture the full duration or frequency of glacial activity (Baker and Carter, 2019; Sinha and Vijayan, 2017). Interpreting climate from these ages requires integration with methods and findings from Chapter 5 (i.e. geomorphology, stratigraphic/superposition relationships, degree of degradation, and cross-cutting landforms). This interdisciplinary approach is limited to providing only relative timing constraints and insights into shifting periods of climatic conditions over time. More specifically, I hope to further distinguish between buried glacial generations, constrain the duration of stable ice conditions, and test whether the stratigraphic transitions observed are regionally unique to Phlegra Montes. This may help answer a research question such as: *Did Phlegra Montes experience multiple generations of glaciations? If so, how many periods and what is the evidence to show this?*

7.4.3 A Methods Review in Mapping & Naming Planetary Landforms

The Mars nomenclature used to describe Martian glacial features, particularly *viscous flow* features (VFFs) has developed incrementally over decades, largely through morphological observations from orbital imagery (Section 3.2.3; **Table 3.2**). Terms like *lobate debris aprons* (LDAs), *lineated valley fill* (LVF), and *concentric crater fill* (CCF) were introduced to describe specific surface expressions of “flow” based on geometry and location context (i.e. slopes, valleys, crater interiors). While these terms can be sometimes perceived as inconsistent or overlapping, they reflect a pragmatic response to the limitations of remote observation and the absence of ground-truth.

By contrast, glacier landforms on Earth are typically named using a process-form framework, where the term conveys both the mechanism of formation (process) and the resulting morphology (form). For example, a *terminal moraine* is a ridge (form) created by sediment deposition at the terminus of a glacier (process). Other examples include *drumlins* (streamlined hills formed by subglacial deformation), *eskers* (sinuous ridges formed by meltwater flow within or beneath glaciers). This naming approach is possible on Earth because processes can be observed or inferred with high confidence from field studies, monitoring, and climate data.

However, Martian glacial nomenclature tends to combine *form* with *context*, intentionally avoiding strong assumptions about origin or genesis. This caution is warranted, as directly inferring process from form on Mars is inherently uncertain due to limited ground truth and its unique environmental conditions. Therefore, imposing Earth-style classifications onto Martian features can be problematic and potentially misleading. Moreover, while there may be opportunity to revise or replace older terms for clarity or consistency, doing so without clear cross-referencing (i.e. renaming CCF) risks fragmenting the literature and undermining decades of accumulated knowledge. Any evolution in naming conventions must be done thoughtfully, preserving the traceability and utility of legacy terminology.

I believe that a *Review of Mars VFFs nomenclature* is necessary to bridge the gap between Mars and Earth glacial science communication and collaboration (i.e. VFF = debris-covered glacier). A more constructive path forward is to supplement the current system with quantitative, process-agnostic descriptors. For example, the Confinement Index (CI) introduced in Chapter 5 offers a reproducible metric or trinary code (i.e. confined, unconfined, semi-confined) for characterizing glacier

morphology across a continuum of glacier morphologies. This supports more standardized mapping and enhances comparability with terrestrial analogues, especially in morphometric and remote sensing contexts. Ultimately, the goal is not to discard the existing nomenclature, but to build a more consistent, interoperable, and reproducible framework for describing ice-related landforms that supports both Mars-specific research and comparative planetology.

7.4.4 Lava-ice interactions

In Chapter 5, I mapped several textures on the surface of VFFs and interpreted potential processes that could be driving the formation of these surface texture facies. Adjacent to some VFFs ($<40^\circ$), on the surface of the plains there seems to be other textures present that I was not able to recognize or confidently map. These textures appear stratigraphically complex, with ambiguous surface boundaries and possibly reworked or embayed relationships with nearby flow features (**Figure 7.1**) I hypothesize that these enigmatic patterns may result from *lava-ice* interactions, which could explain their irregular surface expression and stratigraphic ambiguity.

Lava is documented in Adams Crater within the southern region of Phlegra Montes and is also visible in SHARAD radargrams acquired during this study. Given the Amazonian age of many volcanic units in the region (Tanaka et al., 2014), it is plausible that lava flows postdate and interact with underlying or adjacent ice-rich units. In particular, Hepburn et al. (2020a) described *ghost LDAs* (GLDAs), which are remnant or embayed debris-covered glaciers, preserved beneath or altered by lava flows. Their proposed formation model suggests that lava can overrun, entomb, and partially preserve ice-rich features, leaving behind subtle geomorphic signatures detectable via high-resolution imaging and radar (**Figure 7.1**). I observed similar raised ridges and embayment boundaries around select VFFs in Phlegra Montes, supporting the idea that some features may represent *lava-modified glaciers* rather than purely glacial or volcanic surfaces.

Given Phlegra Montes' combined tectonic, volcanic, and cryospheric history, future work could examine whether cryptic subsurface ice is preserved under lava or if past volcanism disrupted the surface expression and morphology of VFFs. Comparative studies with terrestrial subglacial volcanic systems, such as those in Iceland or British Columbia, Canada (Allen et al., 1982; Björnsson, 2003; Gudmundsson et al., 1997) could help interpret these Martian analogues, particularly in terms of ice preservation, melt dynamics, and lava-ice interaction processes.

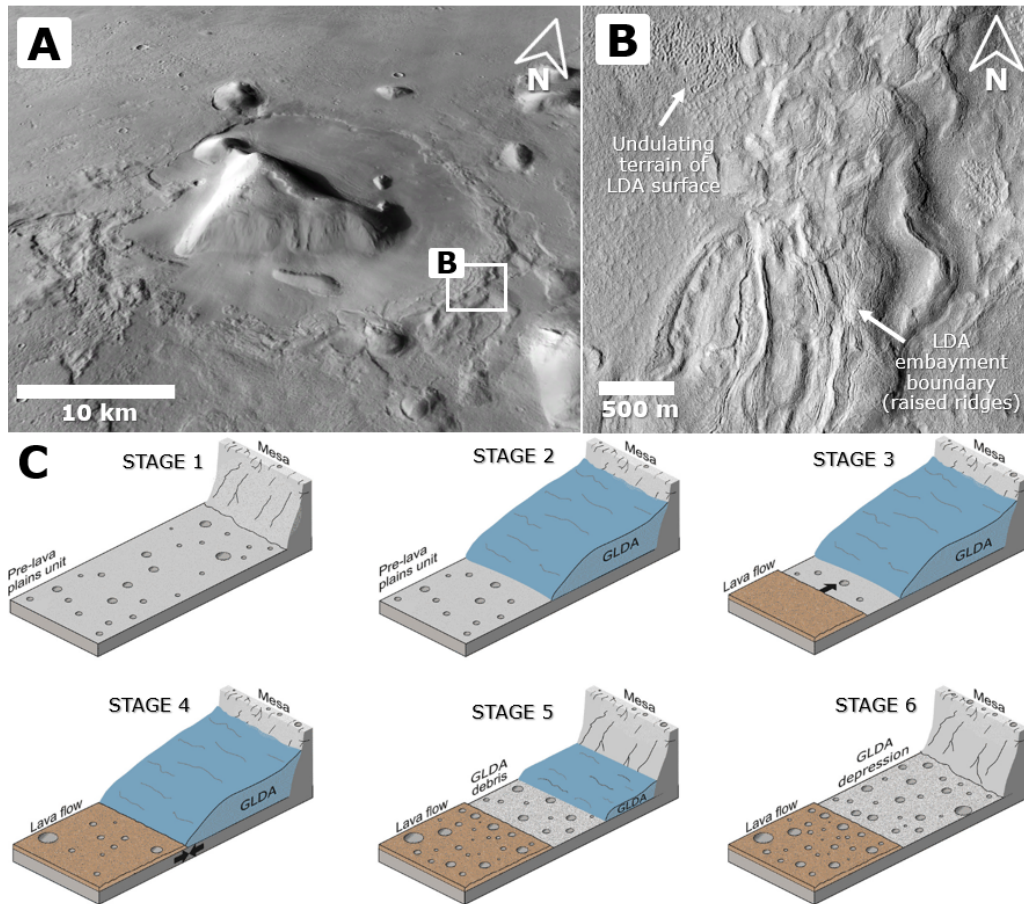


Figure 7.1. (A) Candidate lava-embayed (ghost) LDA at Philgera Montes (32N, 160E). Inset (B) shows HiRISE Image ESP_027767_2145 highlighting the boundary between the LDA and the outside surface where there is a raised ridge. (C) Shows the formation process of a ghost LDA as proposed by Figure 2 in Hepburn et al. (2020a).

7.4.5 Terrestrial Fieldwork on Rock glaciers vs Debris-covered glaciers

The potential for rock glacier and debris-covered glacier as analogues for Mars remains understudied. Future research should investigate whether Martian landforms with lobate or hummocky textures, especially those with minimal radar returns, could represent Martian equivalents of cold-based or inactive rock glaciers or debris-covered glaciers. This poses the question of: *Are rock glaciers and debris-covered glaciers a continuum?* If so, *how does a debris-covered glacier evolve into to rock glacier?* (Anderson et al., 2018; Anderson and Anderson, 2025; Giardino and Vitek, 1988; Johnson, 1980; Meng et al., 2023). Fieldwork and validation in the Canadian Arctic or the Dry Valleys of Antarctica could provide more insight on this work. For now, this is mostly relevant and limited to terrestrial applications where fieldwork can be done. Unfortunately, on Mars, we would need to probe into the icy subsurface of interpreted VFFs with a higher resolution radar or coring instrument in order to see whether they are indeed debris-covered glaciers or rock glaciers.

7.4.6 Drone-based GPR

Building on fieldwork in Yukon in Chapter 6, future research could also include deploying drone-based ground penetrating radar (GPR) to map not only ice wedge networks and active layer thickness in hard-to-reach Arctic terrains, but also on debris-covered glaciers or rock glaciers (Aguilar et al., 2024; Ruols et al., 2023; Tjoelker, 2024; Tjoelker et al., 2024). These systems offer spatial flexibility and resolution that exceed towed GPR setups, serving as analogue platforms for future Mars missions. A drone-based GPR system would alleviate physical labour and automate navigating topographic obstacles, which would also minimize human error in data collection. As an example, such field deployments could also test new radar frequencies and processing pipelines adaptable to airborne or orbital Mars radar missions.

7.4.7 Phlegra Montes as the Next Candidate Human Landing Site

Phlegra Montes is a compelling candidate for future human exploration due to its co-located buried ice, geologic diversity, and favorable mid-latitude position, which offers a balance between accessibility and thermal conditions. As shown in this thesis, numerous LDAs and VFFs in Phlegra contain radar-confirmed subsurface ice and exhibit well-preserved glacial morphologies. These features are not only promising for in-situ resource utilization (ISRU) but also offer valuable targets for reconstructing Mars' recent climate history.

Previous SpaceX landing site workshops (e.g., 2018, 2020) identified several candidate LDAs in Phlegra, though some remain under-imaged. A scoring matrix approach, factoring in ice thickness, latitude, slope, terrain safety, geologic diversity, and ISRU potential, could help prioritize landing zones. This matrix can build on existing selection frameworks (e.g., McEwen et al., 2020) and incorporate updated SHARAD and geomorphic data from this thesis. Beyond ISRU, Phlegra's complex tectonic and glacial history makes it an ideal site for human-led geological investigations. The region contains a variety of surface textures, stratigraphic relationships, and glacial generations that could yield insight into broader questions of Martian geologic and climate evolution.

Following this thesis, I am currently working on a manuscript with Dr. Alfred McEwen to submit to the PSJ Focus Issue on Mars Resources and Science Targets for Human Exploration.

CHAPTER 8

8 Chapters Summary

8.1 Chapters 1-3: Introduction and Literature Review

Chapters 1 to 3 establish the context, aims, and foundational knowledge for this thesis. *Chapter 1* outlines the research goals, which are broadly to characterize buried glacial ice in Phlegra Montes, Mars and evaluate radar methods using analogue fieldwork in the Yukon Territory in the Canadian Arctic. *Chapter 2* describes the main study sites of focus in the thesis, Phlegra Montes and Tombstone Territorial Park, and the multi-method approach, including SHARAD radar analysis, geomorphological mapping, and ground-based geophysics. *Chapter 3* presents a meta-analysis of the cryosphere on Earth and Mars, focusing on glacial and periglacial processes, the evolution of Martian mid-latitude ice, and open questions *surrounding viscous flow features (VFFs) and the latitude dependent mantle (LDM)*.

8.2 Chapter 4: SHARAD in Phlegra Montes

Chapter 4 applies SHARAD radar data to the detection and analysis of buried glaciers in Phlegra Montes. A key innovation introduced here is a two-layer analytical dielectric model that refines traditional radar depth correction by accounting for a debris-lag layer overlying ice. This model significantly improves estimates of total ice thickness and better matches radar reflectors with surface morphology.

Notably, this chapter reports the first radar detections of terraced and hanging glaciers on Mars. These internal reflectors, confirmed through clutter simulation and depth correction, suggest multi-phase glacial activity and ice deposition, likely driven by obliquity-induced climate cycles. The radar results also emphasize the diversity in glacial form and depth across Phlegra Montes, with some reflectors found as deep as ~390 m. Overall, this chapter demonstrates the effectiveness (and limits) of SHARAD in characterizing Martian debris-covered glaciers and establishes baseline targets for future high-resolution radar missions.

8.3 Chapter 5: Geomorphology of Glaciers in Phlegra Montes

Chapter 5 presents a comprehensive geomorphological analysis of 480+ VFFs across Phlegra Montes. Using high-resolution imagery, elevation data, and a new classification tool (i.e. Confinement Index) this chapter moves beyond traditional glacier naming schemes (i.e. LDA/LVF/CCF) to quantify glacier-topography interactions.

The results reveal strong latitudinal and textural trends, with southern VFFs showing exposed, relict surfaces and northern features buried by the latitude dependent mantle (LDM).

Stratigraphic relationships among surface textures (*Lineated, Knobby, Smooth, Undulating, Scalloped* terrains) indicate a multi-generational glacial history, with older VFFs being progressively buried and overprinted by younger LDM. This chapter also confirms a poleward increase in radar-detected ice thickness and preservation, supporting the idea that current Martian mid-latitude glaciers represent a palimpsest of multiple climatic regimes.

8.4 Chapter 6: Geophysical Fieldwork in the Yukon

Chapter 6 shifts focus to Earth, detailing geophysical fieldwork in Tombstone Territorial Park, Yukon, a terrestrial analogue for Mars' icy mid-latitudes. Using multi-frequency ground penetrating radar (GPR), airborne SlimSAR, and capacitively-coupled resistivity (CCR), this chapter evaluates the detectability of shallow ground ice under varying surface conditions.

The field surveys successfully imaged ice wedges, potential massive ice bodies, and permafrost boundaries at depths of 1–5 m. Key findings show that 250 MHz and 500 MHz GPRs provide the best compromise between resolution and penetration depth, crucial knowledge for designing future Martian radar systems. The integration of GPR with SlimSAR also demonstrated how multi-scale radar datasets can resolve both fine and broad subsurface structures, offering a direct analogue to orbital radar studies on Mars.

This chapter highlights the essential role of analogue fieldwork in validating planetary radar methods and guiding future exploration technology.

8.5 Chapter 7: Synthesis

Chapter 7 brings together the results from Mars and Earth to synthesize a broader understanding of the Martian cryosphere and its analogue applications. It confirms that the Phlegra Montes region preserves evidence of multi-generational glaciation, with LDM overlying and modifying earlier VFFs. Radar and geomorphology results converge to support a model of episodic glacial activity influenced by orbital forcing.

The chapter also lays out future research questions and directions, including: **(1)** numerical modeling of glacial flow (e.g., ISSM/IGM) in Phlegra Montes, Mars, **(2)** formalizing landform classification with the Confinement Index, **(3)** investigating rock glaciers vs debris-covered glacier terrestrial analogues, **(4)** understanding lava-ice interactions in Phlegra Montes, Mars, **(5)** expanding radar field methods with drone-based GPR, and **(6)** strengthening the case for Phlegra Montes to be the next candidate human landing site on Mars. This synthesis shows how combining geomorphological interpretation, radar geophysics, and Earth analogue validation can yield powerful insights into planetary surface processes and mission planning.

8.6 Conclusion

This thesis set out to explore how Mars' mid-latitude glacial systems, particularly in *Phlegra Montes, Mars* as well as ground-ice and permafrost in the *Yukon Territory, Canada*, store, preserve, and express near-surface water-ice, and how we can improve their detection and classification using orbital radar, geomorphological analysis, and Earth analogue fieldwork.

These goals were formalized through three core objectives:

1. **Objective #1:** To estimate ice thickness and lag depth in Phlegra Montes using SHARAD radar, and develop improved dielectric modeling to characterize debris-covered glaciers (*Chapter 4*).
2. **Objective #2:** To map, classify, and interpret the geomorphology of viscous flow features (VFFs) in Phlegra Montes, and understand their relationship to the latitude dependent mantle (LDM; *Chapter 5*).
3. **Objective #3:** To validate radar-based ice detection methods using analogue geophysical surveys in the Yukon, and assess the best tools and frequencies for shallow ice characterization (*Chapter 6*).

Each of these objectives was directly addressed. *Objective #1* was met through the development of a novel two-layer dielectric model for SHARAD, accounting for radar signal delay introduced by the debris-lag layer. This advancement improved depth correction and revealed subsurface reflectors with more accurate geometry. It enabled the identification of terraced and hanging glaciers, previously undocumented morphologies that reflect multi-phase glacial evolution. These detections extend the current radar literature and demonstrate that Martian glaciers may preserve complex internal architectures akin to cold-based glaciers on Earth.

Objective #2 was fulfilled through the detailed mapping of over 480 VFFs using high-resolution imagery and the introduction of the Confinement Index, a new metric for classifying glacial landforms based on their degree of topographic enclosure. The results showed clear latitudinal patterns in glacial morphology and surface texture, with southern VFFs displaying relict characteristics and northern features being progressively buried by the LDM. Stratigraphic evidence, such as knobby and lineated terrain overlain by smooth, scalloped units, supported the interpretation of a multi-generational landsystem. The findings clarified the relationship between VFFs and the LDM, suggesting that they are distinct but overlapping systems, rather than products of a single depositional mode.

Objective #3 was addressed and partially fulfilled through field-based geophysical surveys in Tombstone Territorial Park, Yukon, Canada. Ground penetrating radar (GPR), SlimSAR, and resistivity methods revealed subsurface structures such as ice wedges, massive ice, and stratified permafrost at 1–5 m depth. The fieldwork showed that 250–500 MHz GPR frequencies provide the best balance of resolution and penetration, information directly transferable to future Mars radar design, including proposed missions like I-MIM. These analogue results validated radar methods and confirmed that shallow ground ice is detectable under varied surface and material conditions, especially when guided by optimized dielectric models.

Together, these findings present a robust, multi-method characterization of Phlegra Montes as a glaciologically rich and complex region, one that preserves a detailed record of Mars' climate history. This thesis makes original contributions to three key areas: *detection*, *classification*, and *validation*. *Detection* refers to a novel radar analytical model and new subglacial reflectors mapped at higher fidelity using SHARAD on Phlegra Montes, Mars. *Classification* is illustrated

through the new mapping methods, quantification (i.e. Confinement Index), and surface-texture mapping and stratigraphy revealing glacial cycles at Phlegra Montes, Mars giving insights for the northern midlatitudes of Mars. *Validation* refers to field-tested geophysical tools that strengthen the case for future Mars missions highlighting the Yukon Territory, Canada.

Beyond answering the initial research questions, this work opens new lines of inquiry. Chapter 7 outlines some general future directions from modeling glacial flow dynamics, formalizing planetary landform classification schemes, studying lava-ice interactions, to conducting terrestrial fieldwork on rock and debris-covered glaciers perhaps by using drone-based GPR for field analogue surveys. Each of these next steps builds directly on the frameworks established here.

Importantly, this thesis shows that studying Mars' cryosphere cannot rely on a single method or dataset. It requires a combination of interdisciplinary datasets such as orbital data, geomorphic theory, field analogues, and technical innovation. The icy landsystems of Phlegra Montes are not static remnants of the past, they are dynamic systems whose formation, degradation and/or preservation are governed by recurring climate cycles, topographic controls, and surface processes that are familiar to us from Earth. As planetary exploration shifts toward in-situ science, resource mapping, and potential human missions, the detection and understanding of shallow ice (1-5 m) will be essential. This work helps define where that ice is, how it formed, how it persists, and how we can best detect it. It reaffirms the value of mid-latitude ice as a scientific target, a climate archive, and a practical asset for future exploration missions.

This thesis has contributed new discoveries, developed new tools, validated detection methods, and laid out a clear vision for what may come next. It bridges not only two planets, but interdisciplinary techniques through the investigation of the cryosphere on both Earth and Mars as one system of landforms and processes that evolve through time. In doing so, it takes a significant step forward in understanding not only Mars' frozen past, but also its possible future by using our own planet, Earth, as a reference point.

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Appendix

Table A1. SHARAD tracks with positive subsurface reflectors for all study sites at Phlegra Montes. Respective labels (second subcolumn) representing individual radar tracks with the highlighted track denoted with (*). Observation numbers with (*) indicate previously studied tracks. ΔMOLA thickness is the difference between MOLA high and low and is independent of SHARAD

Site	Observation #	Subsurface Two-way time (ms)	Surface Two-way time (ms)	TWT	ΔTwo-way time (s)	MOLA High (m)	MOLA Low (m)	ΔMOLA thickness (m)	ε _{bulk}	Mean	Std Dev ±	RMS E	RMSE Mean	ε _{ice3.15}	ε _{lag8}	ε _{lag5.5}	Lag thickness (8) (m)	Lag thickness (5.5) (m)	Ice thickness (8) (m)	Ice thickness (5.5) (m)
1	A 13176_01*	916	866.5	49.5	#### ##	-2979	-3394	415	3.2	3.34	0.131	2.00E-02	0.125876516	3.15	8	5.5	1.29	1.44	413.71	413.56
	B 31294_01*	917.4	866.1	51.3	#### ##	-2854	-3260	406	3.59			6.24E-02		3.15	8	5.5	10.63	11.87	395.37	394.13
	C 36608_01*	916.1	872.2	43.9	#### ##	-2910	-3271	361	3.33			2.28E-04		3.15	8	5.5	3.86	4.32	357.14	356.68
	D 45628_01*	914.3	866.6	47.7	#### ##	-2930	-3329	399	3.22			1.61E-02		3.15	8	5.5	1.6	1.78	397.4	397.22
	E 45984_02*	917.7	867.5	50.2	#### ##	-2855	-3265	410	3.37			9.37E-04		3.15	8	5.5	5.5	6.15	404.5	403.85
	F 46551_01*	914.9	867.1	47.8	#### ##	-2933	-3315	382	3.52			3.26E-02		3.15	8	5.5	8.48	9.47	373.52	372.53
	G 48331_01	913	871.4	41.6	#### ##	-2995	-3325	330	3.58			5.43E-02		3.15	8	5.5	8.32	9.3	321.68	320.7
	H 56533_02	911	872.6	38.4	#### ##	-2985	-3302	317	3.3			1.67E-03		3.15	8	5.5	2.91	3.25	314.09	313.75
	I 59131_02	912.9	869.3	43.6	#### ##	-2981	-3342	361	3.28			3.65E-03		3.15	8	5.5	2.89	3.23	358.11	357.77
	J 59909_01	911.6	872.4	39.2	#### ##	-3005	-3329	324	3.29			2.39E-03		3.15	8	5.5	2.81	3.14	321.19	320.86
	K 64155_01	910.5	872.2	38.3	#### ##	-3004	-3323	319	3.24			9.81E-03		3.15	8	5.5	1.81	2.02	317.19	316.98
	L 67281_01	911.8	874.9	36.9	#### ##	-3068	-3374	306	3.27			4.98E-03		3.15	8	5.5	2.26	2.53	303.74	303.47
	M 69432_01	913.6	869.5	44.1	#### ##	-2983	-3351	368	3.23			1.24E-02		3.15	8	5.5	1.82	2.03	366.18	365.97
	N 70210_01	915.6	864.5	51.1	#### ##	-2911	-3329	418	3.36			4.05E-04		3.15	8	5.5	5.35	5.98	412.65	412.02
2	A 09418_03	891.9	870.1	21.8	#### ##	-2930	-3106	176	3.45	3.35	0.07	1.05E-02	0.078950954	3.15	8	5.5	3.18	3.55	172.82	172.45
	B 11897_01	885.4	855.3	30.1	#### ##	-2869	-3115	246	3.37			3.54E-04		3.15	8	5.5	3.24	3.61	242.76	242.39
	C 38296_01	890.1	863.4	26.7	#### ##	-2865	-3086	221	3.28			4.30E-03		3.15	8	5.5	1.8	2.01	219.2	218.99
	D 48041_01	893.2	867.5	25.7	#### ##	-2893	-3104	211	3.34			1.38E-04		3.15	8	5.5	2.39	2.67	208.61	208.33
	E 67954_01	898.9	863.7	35.2	#### ##	-2907	-3197	290	3.31			1.21E-03		3.15	8	5.5	2.89	3.23	287.11	286.77
	F 68020_01*	889.3	864.8	24.5	#### ##	-2883	-3086	203	3.28			5.24E-03		3.15	8	5.5	1.57	1.75	201.43	201.25

	G	69221_0 1	888	867.8	20.2	#### ##	-2918	-3087	169	3.21			1.83 E-02		3.15	8	5.5	0.66	0.74	168.34	168.26
	H	69920_0 2	887.3	855.1	32.2	#### ##	-2910	-3178	268	3.25			1.03 E-02		3.15	8	5.5	1.6	1.78	266.4	266.22
	I	74825_0 1	881.9	862.6	19.3	#### ##	-2920	-3080	160	3.27			5.76 E-03		3.15	8	5.5	1.2	1.34	158.8	158.66
3	A	59276_0 1	-	-			-	-	-												
	B	60265_0 2	-	-			-	-	-												
	C	61109_0 1	935.5	900.1	35.4	#### ##	-3325	-3619	294	3.26	3.24	0.03	4.20 E-04	0.0204989 84	3.15	8	5.5	2	2.23	292	291.77
	D *	66806_0 1	928.6	894.5	34.1	#### ##	-3339	-3624	285	3.22			4.20 E-04		3.15	8	5.5	1.23	1.38	283.77	283.62
	E	69366_0 2	-	-			-	-	-												
4	A	67149_0 1	-	-			-	-	-												
	B *	69643_0 2	-	-			-	-	-												
	C	72280_0 1	-	-			-	-	-												
5	A	12609_0 1	872	848.9	23.1	#### ##	-2540	-2730	190	3.33	3.28	0.07	1.78 E-03	0.0682314 13	3.15	8	5.5	2.02	2.25	187.98	187.75
	B	35408_0 1	871.8	849.3	22.5	#### ##	-2529	-2716	187	3.26			6.91 E-04		3.15	8	5.5	1.22	1.36	185.78	185.64
	C	38230_0 1	871.6	846.2	25.4	#### ##	-2556	-2766	210	3.29			6.40 E-05		3.15	8	5.5	1.8	2.01	208.2	207.99
	D	49043_0 1	869.1	846.1	23	#### ##	-2537	-2723	186	3.44			2.46 E-02		3.15	8	5.5	3.23	3.61	182.77	182.39
	E *	62731_0 1	868.8	844.3	24.5	#### ##	-2596	-2800	204	3.25			1.47 E-03		3.15	8	5.5	1.18	1.32	202.82	202.68
	F	65421_0 3	871.3	848.5	22.8	#### ##	-2636	-2827	191	3.21			6.00 E-03		3.15	8	5.5	0.65	0.73	190.35	190.27
	G	69155_0 1	874.9	856.1	18.8	#### ##	-2608	-2764	156	3.27			2.52 E-04		3.15	8	5.5	1.11	1.24	154.89	154.76
	H	74258_0 1	872.6	848.5	24.1	#### ##	-2507	-2708	201	3.23			2.40 E-03		3.15	8	5.5	1.03	1.16	199.97	199.84
6	A	04711_0 1	871	834.8	36.2	#### ##	-2369	-2669	300	3.28	3.24	0.02	1.33 E-03	0.0219536 88	3.15	8	5.5	2.29	2.56	297.71	297.44
	B	12609_0 1	869.1	839.3	29.8	#### ##	-2321	-2570	249	3.22			2.89 E-04		3.15	8	5.5	1.1	1.23	247.9	247.77
	C *	46762_0 1	872.5	833.3	39.2	#### ##	-2466	-2793	327	3.23			3.93 E-05		3.15	8	5.5	1.66	1.85	325.34	325.15
	D	57021_0 1	871.3	839.8	31.5	#### ##	-2446	-2709	263	3.23			1.43 E-04		3.15	8	5.5	1.24	1.39	261.76	261.61
	E	62731_0 1	869.5	838	31.5	#### ##	-2300	-2563	263	3.23			1.43 E-04		3.15	8	5.5	1.24	1.39	261.76	261.61
	F	64254_0 2	871.9	835.3	36.6	#### ##	-2505	-2811	306	3.22			4.33 E-04		3.15	8	5.5	1.28	1.43	304.72	304.57
	G	65421_0 3	869.5	836.1	33.4	#### ##	-2357	-2634	277	3.27			9.98 E-04		3.15	8	5.5	2.04	2.28	274.96	274.72
7	A	09418_0 3	887.5	850.7	36.8	#### ##	-2651	-2958	307	3.23	3.23	0.04	1.95 E-05	0.0370882 71	3.15	8	5.5	1.55	1.73	305.45	305.27
	B	12464_0 1	884	849.6	34.4	#### ##	-2652	-2942	290	3.17			3.92 E-03		3.15	8	5.5	0.28	0.32	289.72	289.68

	C	13031_0 1	876.7	846.2	30.5	#### ##	-2529	-2783	254	3.24			2.47 E-04		3.15	8	5.5	1.45	1.62	252.55	252.38
	D	38296_0 1	880.2	848.6	31.6	#### ##	-2731	-2995	264	3.22			2.39 E-05		3.15	8	5.5	1.18	1.32	262.82	262.68
	E	46406_0 1	882.9	849.4	33.5	#### ##	-2670	-2949	279	3.24			2.35 E-04		3.15	8	5.5	1.59	1.78	277.41	277.22
	F *	48041_0 1	883.3	849.2	34.1	#### ##	-2687	-2974	287	3.18			2.73 E-03		3.15	8	5.5	0.46	0.52	286.54	286.48
	G	56599_0 1	878.3	851.8	26.5	#### ##	-2728	-2949	221	3.24			4.31 E-05		3.15	8	5.5	1.14	1.28	219.86	219.72
	H	64788_0 1	873.4	845.6	27.8	#### ##	-2537	-2770	233	3.2			6.51 E-04		3.15	8	5.5	0.75	0.84	232.25	232.16
	I	66318_0 2	873.6	851.3	22.3	#### ##	-2603	-2791	188	3.17			3.94 E-03		3.15	8	5.5	0.18	0.2	187.82	187.8
	J	67954_0 1	883.6	851.4	32.2	#### ##	-2584	-2851	267	3.27			1.93 E-03		3.15	8	5.5	1.98	2.21	265.02	264.79
	K	68020_0 1	877.1	850.4	26.7	#### ##	-2737	-2961	224	3.2			1.01 E-03		3.15	8	5.5	0.64	0.71	223.36	223.29
	L	69221_0 1	878	850.3	27.7	#### ##	-2734	-2964	230	3.26			1.22 E-03		3.15	8	5.5	1.58	1.77	228.42	228.23
	M	69498_0 1	874.8	851.8	23	#### ##	-2576	-2766	190	3.3			4.70 E-03		3.15	8	5.5	1.69	1.89	188.31	188.11
	N	69920_0 2	878.5	847.2	31.3	#### ##	-2704	-2964	260	3.26			1.04 E-03		3.15	8	5.5	1.75	1.95	258.25	258.05
	O	70777_0 1	880.1	847.9	32.2	#### ##	-2728	-2997	269	3.22			2.10 E-05		3.15	8	5.5	1.21	1.35	267.79	267.65
	P	72201_0 1	881.5	852.7	28.8	#### ##	-2690	-2930	240	3.24			1.31 E-04		3.15	8	5.5	1.31	1.47	238.69	238.53
	Q	73691_0 1	873.7	850.1	23.6	#### ##	-2582	-2780	198	3.2			1.03 E-03		3.15	8	5.5	0.56	0.63	197.44	197.37
	R	74825_0 1	877	848.3	28.7	#### ##	-2747	-2985	238	3.27			1.87 E-03		3.15	8	5.5	1.76	1.96	236.24	236.04
8	A	12207_0 1	933.4	911.3	22.1	#### ##	-3481	-3663	182	3.32	3.28	0.08	1.28 E-03	0.0720173 87	3.15	8	5.5	1.84	2.06	180.16	179.94
	B	34050_0 1	935	916	19	#### ##	-3545	-3704	159	3.21			4.75 E-03		3.15	8	5.5	0.61	0.68	158.39	158.32
	C	35540_0 1	933	920.9	12.1	#### ##	-3625	-3723	98	3.43			2.20 E-02		3.15	8	5.5	1.64	1.84	96.36	96.16
	D	37129_0 1	936.2	919.5	16.7	#### ##	-3484	-3621	137	3.34			3.78 E-03		3.15	8	5.5	1.6	1.78	135.4	135.22
	E	46149_0 1	935.3	917.1	18.2	#### ##	-3422	-3572	150	3.31			9.36 E-04		3.15	8	5.5	1.47	1.64	148.53	148.36
	F	59263_0 2	933.5	916.6	16.9	#### ##	-3557	-3698	141	3.23			2.45 E-03		3.15	8	5.5	0.71	0.79	140.29	140.21
	G	60252_0 2	933.6	914.5	19.1	#### ##	-3542	-3701	159	3.25			1.23 E-03		3.15	8	5.5	0.93	1.04	158.07	157.96
	H	60918_0 2	935.1	912.8	22.3	#### ##	-3475	-3663	188	3.17			1.35 E-02		3.15	8	5.5	0.18	0.2	187.82	187.8
	I	66450_0 2	936.1	911.6	24.5	#### ##	-3447	-3649	202	3.31			7.88 E-04		3.15	8	5.5	1.95	2.18	200.05	199.82
	J*	67611_0 1	933.4	914.9	18.5	#### ##	-3547	-3701	154	3.25			1.21 E-03		3.15	8	5.5	0.91	1.01	153.09	152.99
9	A *	04711_0 1	930.4	887.7	42.7	#### ##	-3353	-3709	356	3.24	3.23	0.03	6.40 E-05	0.0312658 6	3.15	8	5.5	1.88	2.1	354.12	353.9
	B	12609_0 1	927.2	889.5	37.7	#### ##	-3319	-3634	315	3.22			3.71 E-05		3.15	8	5.5	1.4	1.56	313.6	313.44

	C	46762_0 1	933.5	888.6	44.9	#### ##	-3222	-3597	375	3.23			1.12 E-05		3.15	8	5.5	1.73	1.93	373.27	373.07
	D	57021_0 1	934.8	889.8	45	#### ##	-3184	-3562	378	3.19			1.62 E-03		3.15	8	5.5	0.89	1	377.11	377
	E	62731_0 1	938.8	887.2	51.6	#### ##	-3106	-3533	427	3.29			3.22 E-03		3.15	8	5.5	3.51	3.92	423.49	423.08
	F	65421_0 3	939.6	893.8	45.8	#### ##	-3173	-3557	384	3.2			7.96 E-04		3.15	8	5.5	1.19	1.33	382.81	382.67
	G	69155_0 1	938.8	889.3	49.5	#### ##	-3177	-3589	412	3.25			3.57 E-04		3.15	8	5.5	2.45	2.74	409.55	409.26
	H	71845_0 1	930.1	887.5	42.6	#### ##	-3305	-3662	357	3.2			6.33 E-04		3.15	8	5.5	1.17	1.31	355.83	355.69
	I	72346_0 1	929.1	888.5	40.6	#### ##	-3392	-3728	336	3.29			3.16 E-03		3.15	8	5.5	2.75	3.07	333.25	332.93
	J	73058_0 1	930.8	888.1	42.7	#### ##	-3210	-3568	358	3.2			7.88 E-04		3.15	8	5.5	1.11	1.24	356.89	356.76
	K	74258_0 1	935.3	889.6	45.7	#### ##	-3149	-3531	382	3.22			7.62 E-05		3.15	8	5.5	1.63	1.82	380.37	380.18

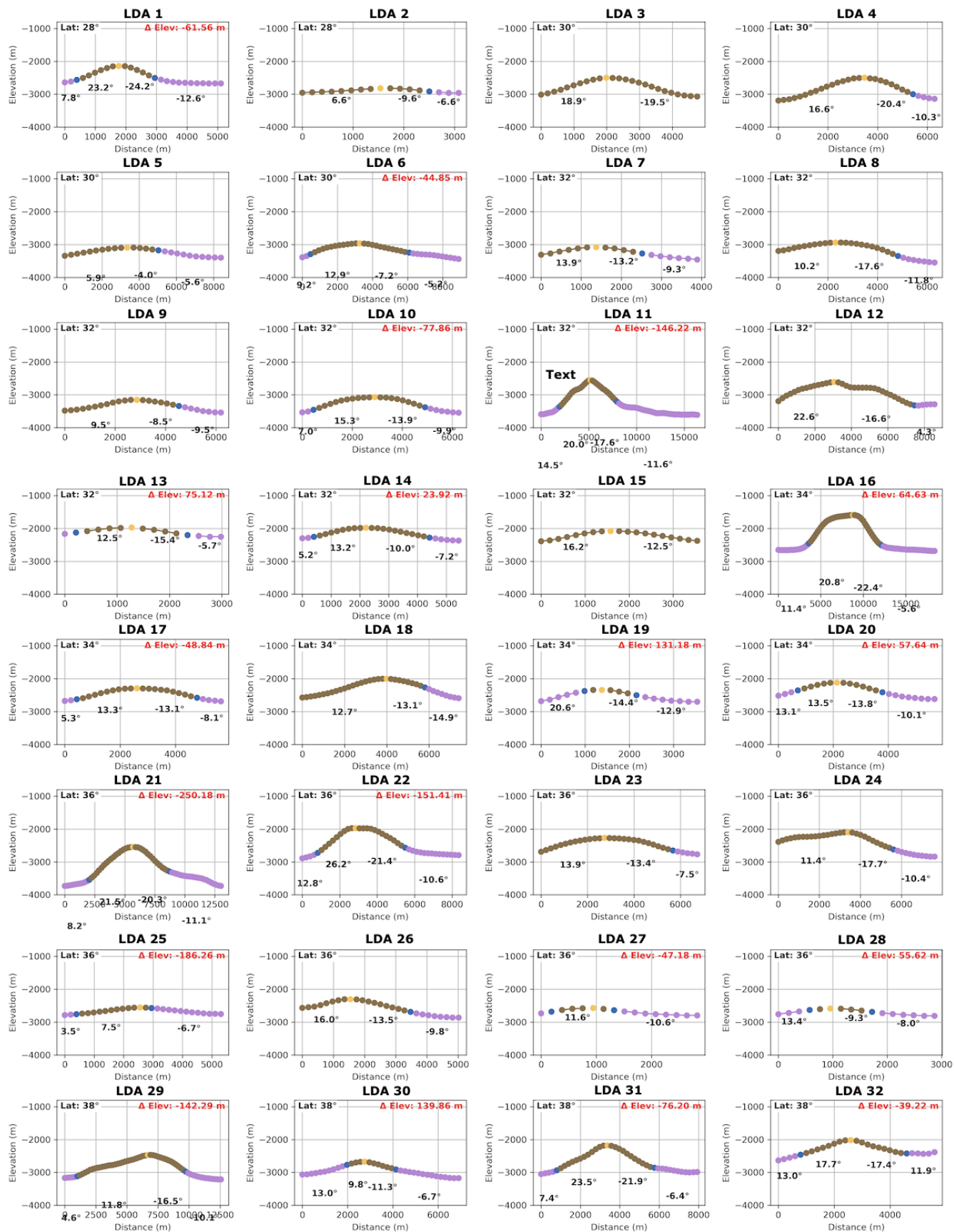
Table A2. List of mapped SHARAD reflectors with positive detections for three LDAs outlined in Section 4.3.2.2.

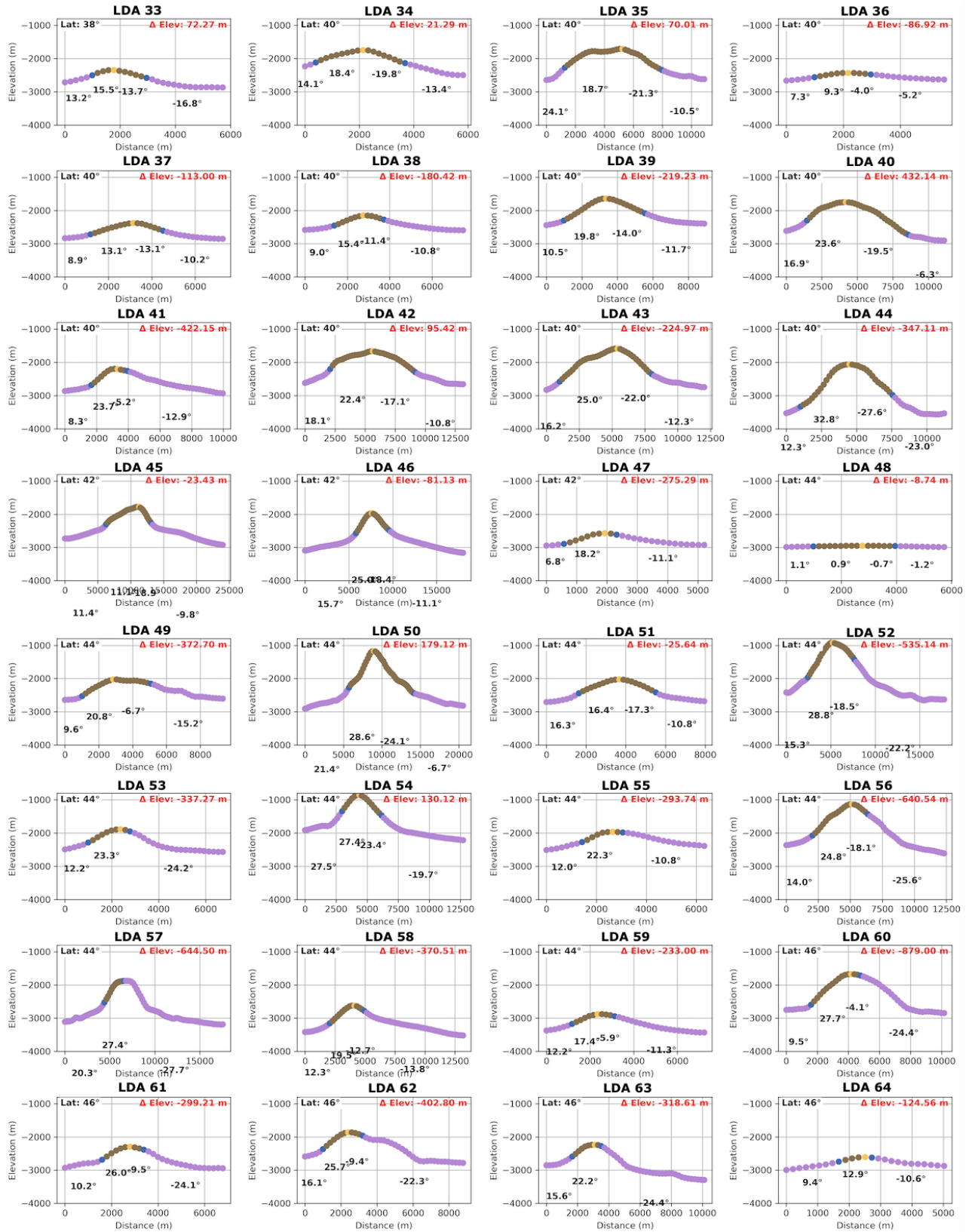
	Observation #		Subsurface Two-way time (ms)	Surface Two-way time (ms)	Δ Two-way time (s)	MOLA High (m)	MOLA Low (m)	Δ MOLA thickness (m)	ϵ_b	Lag thickness (m)	Ice thickness (m)
LDA #50	A	09418_03	887.50	850.70	3.68E-06	-2651.00	-2958.00	307.00	3.23	6.77	300.23
	B	12464_01	884.00	849.60	3.44E-06	-2652.00	-2942.00	290.00	3.17	1.23	288.77
	C	13031_01	876.70	846.20	3.05E-06	-2529.00	-2783.00	254.00	3.24	6.35	247.65
	D	38296_01	880.20	848.60	3.16E-06	-2731.00	-2995.00	264.00	3.22	5.17	258.83
	E	46406_01	882.90	849.40	3.35E-06	-2670.00	-2949.00	279.00	3.24	6.95	272.05
	F*	48041_01	883.30	849.20	3.41E-06	-2687.00	-2974.00	287.00	3.18	2.02	284.98
	G	56599_01	878.30	851.80	2.65E-06	-2728.00	-2949.00	221.00	3.24	5.00	216.00
	H	64788_01	873.40	845.60	2.78E-06	-2537.00	-2770.00	233.00	3.20	3.29	229.71
	I	66318_02	873.60	851.30	2.23E-06	-2603.00	-2791.00	188.00	3.17	0.79	187.21
	J	67954_01	883.60	851.40	3.22E-06	-2584.00	-2851.00	267.00	3.27	8.66	258.34
	K	68020_01	877.10	850.40	2.67E-06	-2737.00	-2961.00	224.00	3.20	2.79	221.21
	L	69221_01	878.00	850.30	2.77E-06	-2734.00	-2964.00	230.00	3.26	6.92	223.08
	M	69498_01	874.80	851.80	2.30E-06	-2576.00	-2766.00	190.00	3.30	7.39	182.61
	N	69920_02	878.50	847.20	3.13E-06	-2704.00	-2964.00	260.00	3.26	7.64	252.36
	O	70777_01	880.10	847.90	3.22E-06	-2728.00	-2997.00	269.00	3.22	5.29	263.71
	P	72201_01	881.50	852.70	2.88E-06	-2690.00	-2930.00	240.00	3.24	5.73	234.27
	Q	73691_01	873.70	850.10	2.36E-06	-2582.00	-2780.00	198.00	3.20	2.45	195.55
	R	74825_01	877.00	848.30	2.87E-06	-2747.00	-2985.00	238.00	3.27	7.68	230.32
LDA #68	A	1288601	931.6	901.3	3.03E-06	-3385.00	-3637.00	252.00	3.25	1.57	250.43
	B	6928702	929.5	901.1	2.84E-06	-3461.00	-3693.00	232.00	3.37	3.09	228.91
	C	6464301	925.6	903.9	2.17E-06	-3494.00	-3669.00	175.00	3.46	3.323	171.76
	D	1231901	923.1	902.7	2.04E-06	-3344.00	-3505.00	161.00	3.61	4.39	156.60
	E	3787401	929.3	904.9	2.44E-06	-3355.00	-3557.00	202.00	3.28	1.62	200.37

LDA #74	A*	04711_01	930.40	887.70	4.27E-06	-3353.00	-3709.00	356.00	3.24	8.22	347.78
	B	12609_01	927.20	889.50	3.77E-06	-3319.00	-3634.00	315.00	3.22	6.10	308.90
	C	46762_01	933.50	888.60	4.49E-06	-3222.00	-3597.00	375.00	3.23	7.54	367.46
	D	57021_01	934.80	889.80	4.50E-06	-3184.00	-3562.00	378.00	3.19	3.91	374.09
	E	62731_01	938.80	887.20	5.16E-06	-3106.00	-3533.00	427.00	3.29	15.33	411.67
	F	65421_03	939.60	893.80	4.58E-06	-3173.00	-3557.00	384.00	3.20	5.19	378.81
	G	69155_01	938.80	889.30	4.95E-06	-3177.00	-3589.00	412.00	3.25	10.70	401.30
	H	71845_01	930.10	887.50	4.26E-06	-3305.00	-3662.00	357.00	3.20	5.11	351.89
	I	72346_01	929.10	888.50	4.06E-06	-3392.00	-3728.00	336.00	3.29	12.02	323.98
	J	73058_01	930.80	888.10	4.27E-06	-3210.00	-3568.00	358.00	3.20	4.85	353.15
	K	74258_01	935.30	889.60	4.57E-06	-3149.00	-3531.00	382.00	3.22	7.13	374.87

Table A3. List of mean aspect values of textures per latitude bin. The first value represents the mean and the second value represents the uncertainty.

Texture	Lat 28	Lat 30	Lat 32	Lat 34	Lat 36	Lat 38	Lat 40	Lat 42	Lat 44	Lat 46	Lat 48	Lat 50
Massif	187.86° ± 2.30°	219.24° ± 2.39°	188.34° ± 2.12°	71.55° ± 2.50°	290.52° ± 1.82°	196.47° ± 2.00°	273.31° ± 2.09°	113.94° ± 1.76°	223.66° ± 1.55°	188.88° ± 1.30°	278.16° ± 2.17°	37.41° ± 1.63°
Knobby	62.13° ± 1.82°	336.20° ± 2.08°	124.99° ± 1.75°	109.50° ± 2.28°	3.02° ± 1.29°	129.82° ± 0.82°	196.31° ± 1.35°	-	215.89° ± 0.55°	-	-	-
Lineated	-	50.73° ± 0.71°	21.30° ± 0.88°	8.10° ± 0.79°	97.95° ± 1.66°	355.36° ± 1.37°	41.89° ± 1.17°	75.69° ± 1.00°	112.94° ± 0.82°	-	-	-
Smooth	-	-	-	-	355.04° ± 1.69°	276.21° ± 1.34°	16.67° ± 1.93°	303.02° ± 1.10°	311.86° ± 1.47°	299.71° ± 1.57°	295.57° ± 1.93°	301.64° ± 1.61°
Undulating	-	-	-	-	-	55.48° ± 2.36°	322.34° ± 1.37°	98.11° ± 1.76°	89.66° ± 1.01°	90.23° ± 1.30°	171.94° ± 1.05°	-
Scallop	-	-	-	-	-	-	-	-	11.90° ± 1.51°	166.87° ± 1.28°	193.95° ± 2.62°	167.63° ± 2.31°





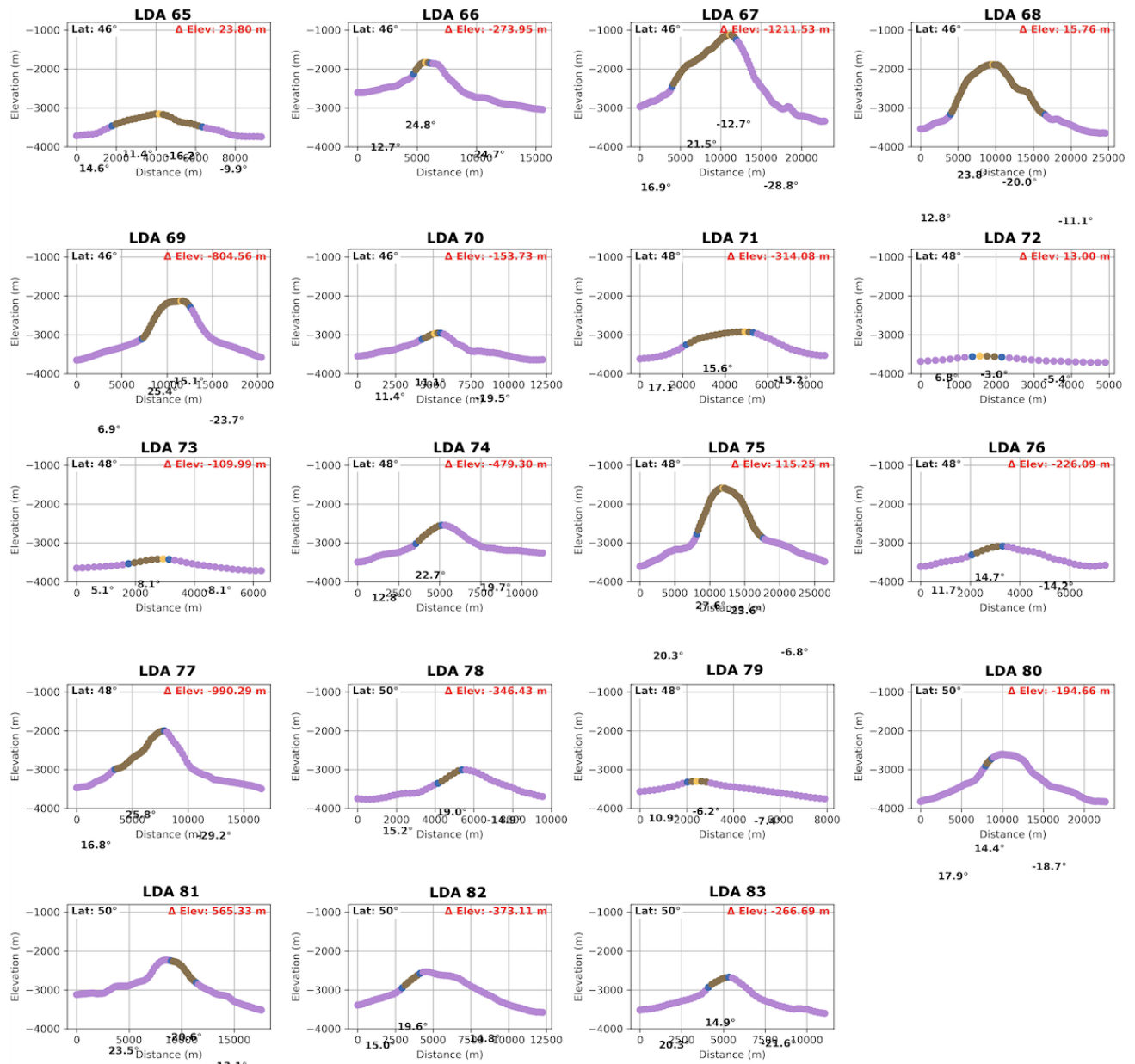
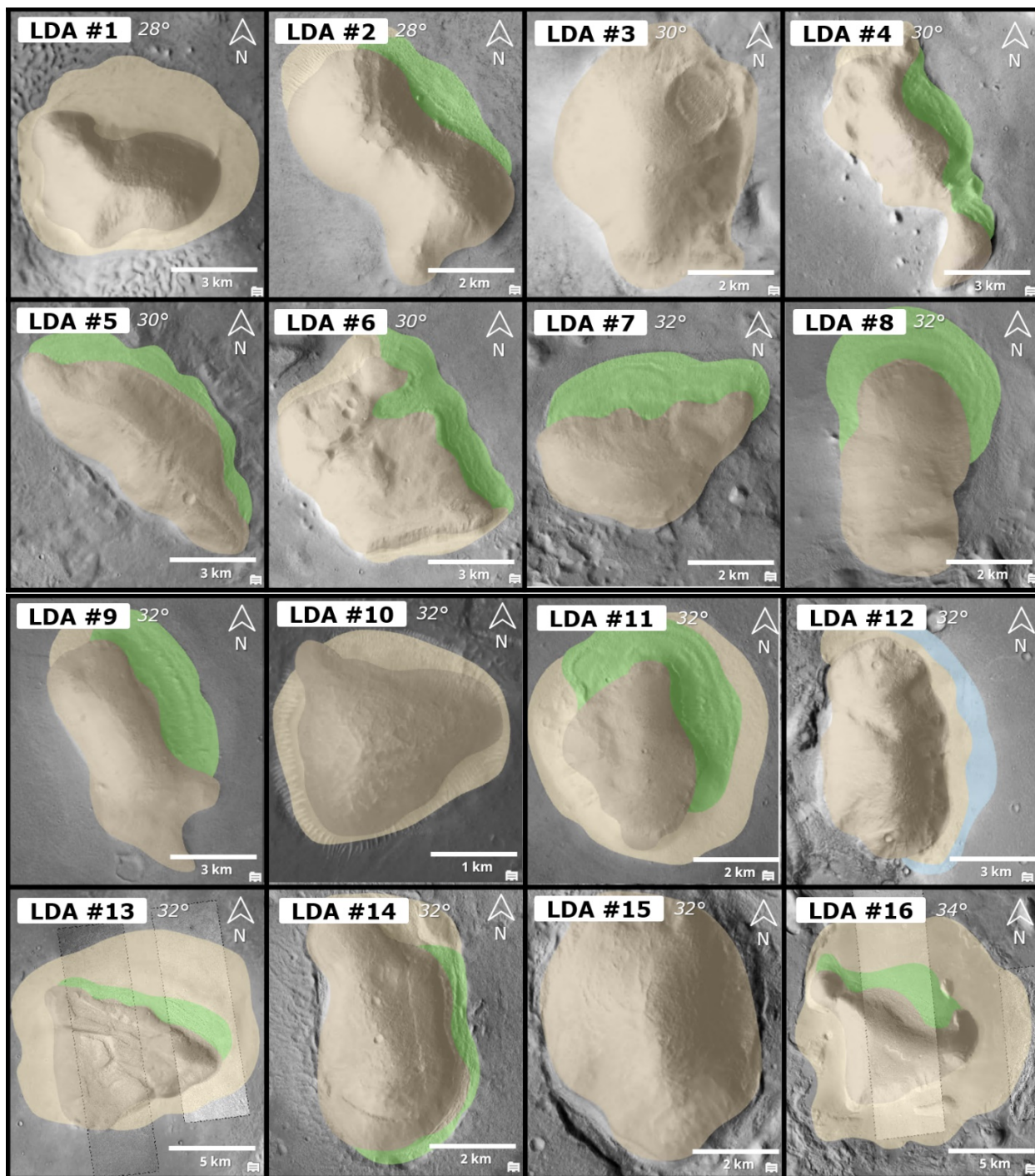
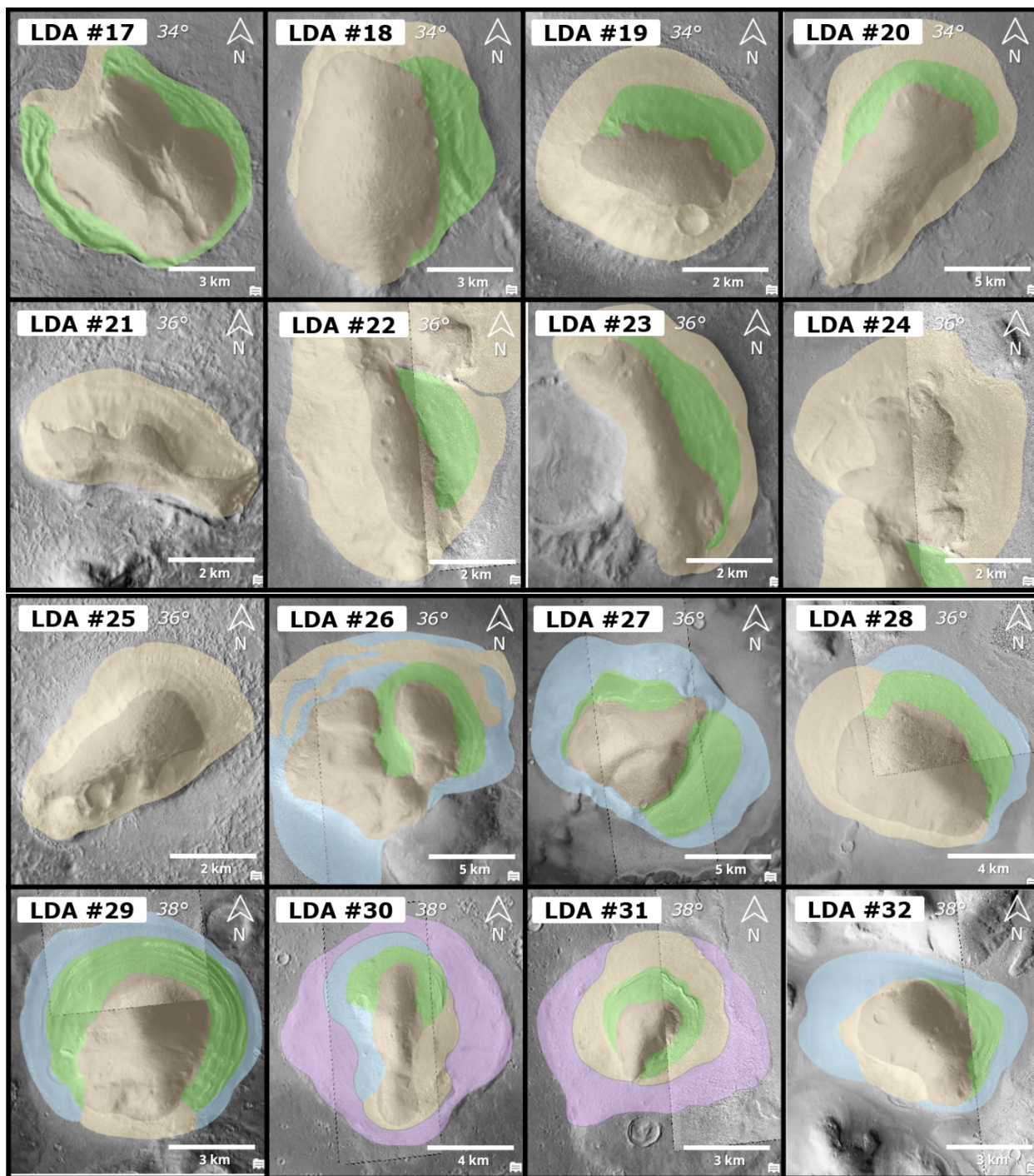
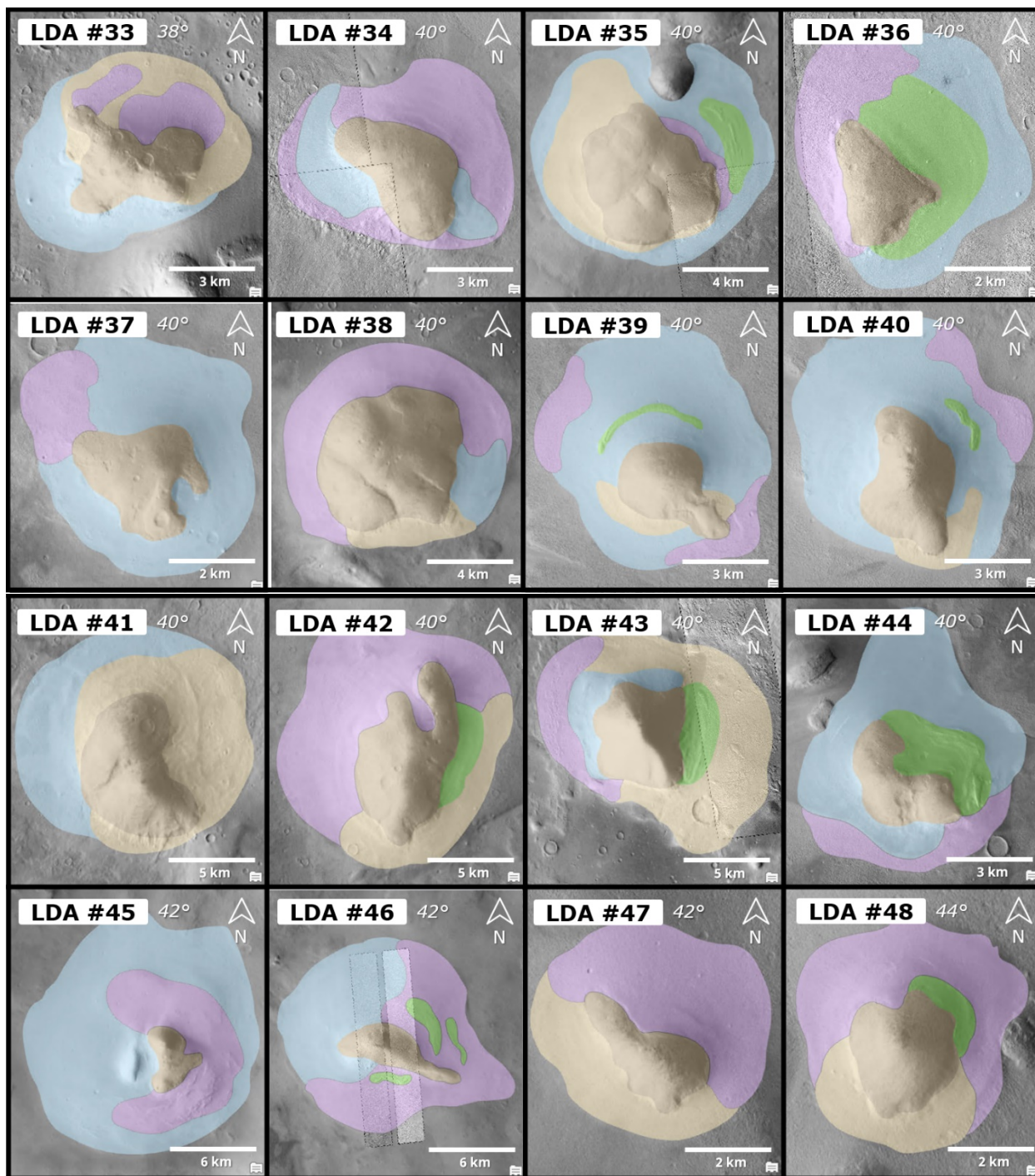
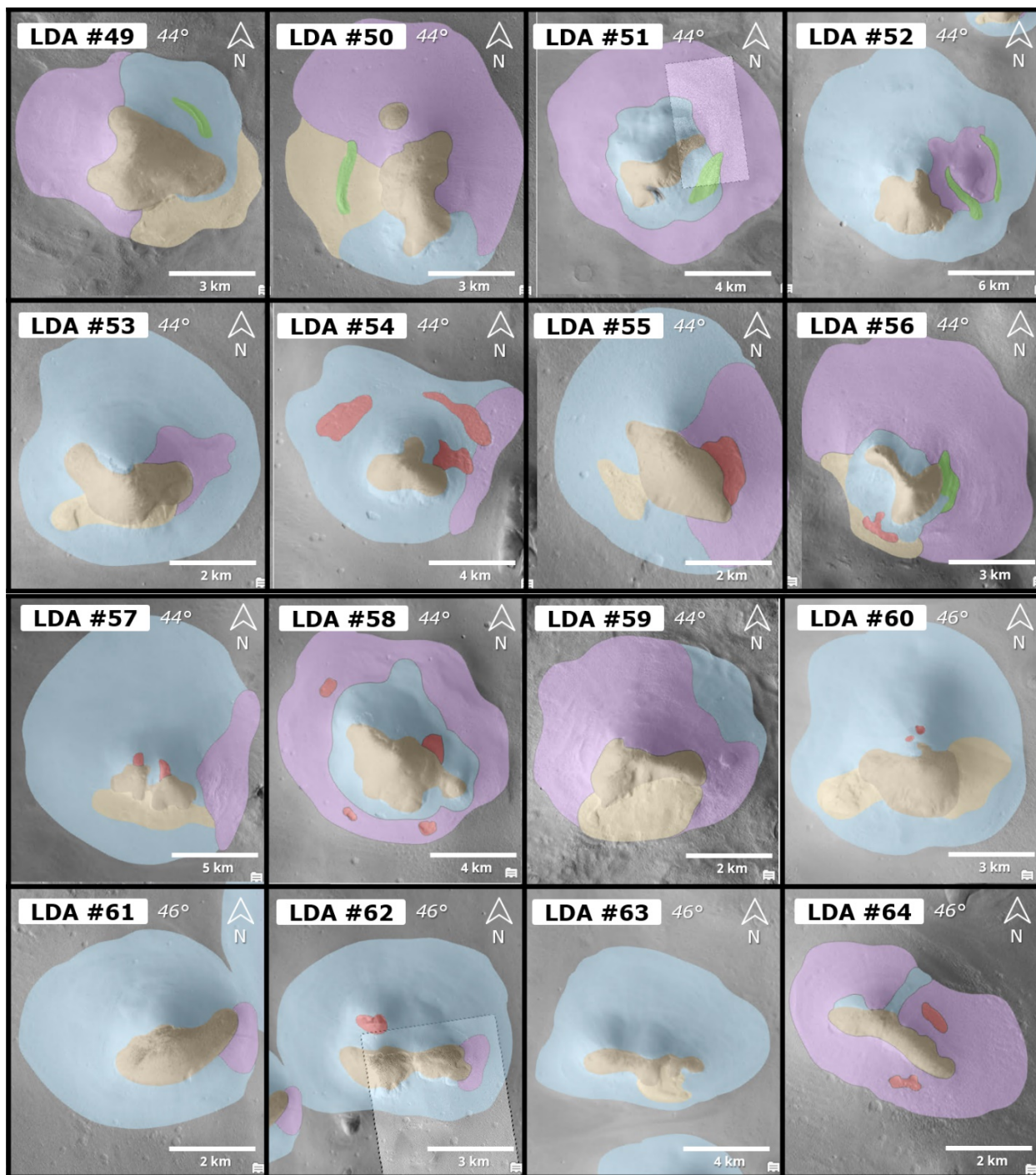


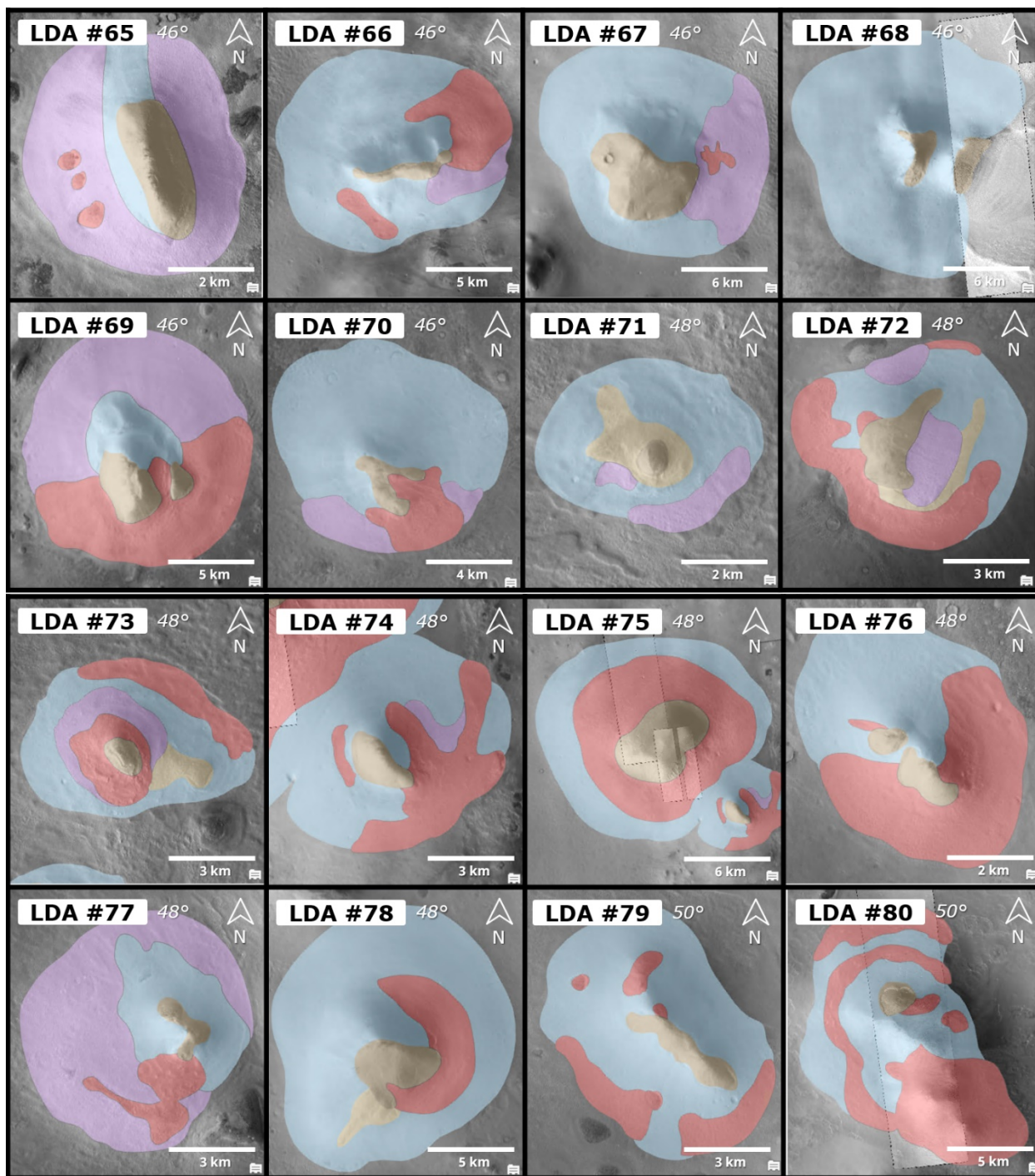
Figure A1. Topographic profiles of LDAs per latitude bin. Segments are colour coded as outlined in Section 2.5.1 where the purple dots represent the LDA apron, the blue dots represent the transition from LDA to massif. The brown dots represent the massif. The yellow dots represent the highest point in the LDA. The slope values for each flank is outlined in blue.











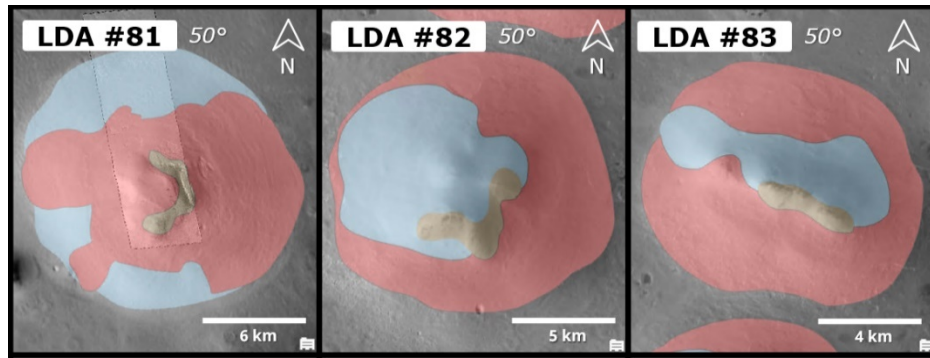


Figure A2. Mapped shapefile boundaries of textures present in each LDA (latitude groups matching Figure A1). The colour legend is as follows: Massif = brown, Knobby = yellow, Lineated = green, Smooth = blue, Undulating = purple, Scalloped = red. Shapefiles are semi transparent showing the CTX Mosaic v1 as the basemap used in the background.

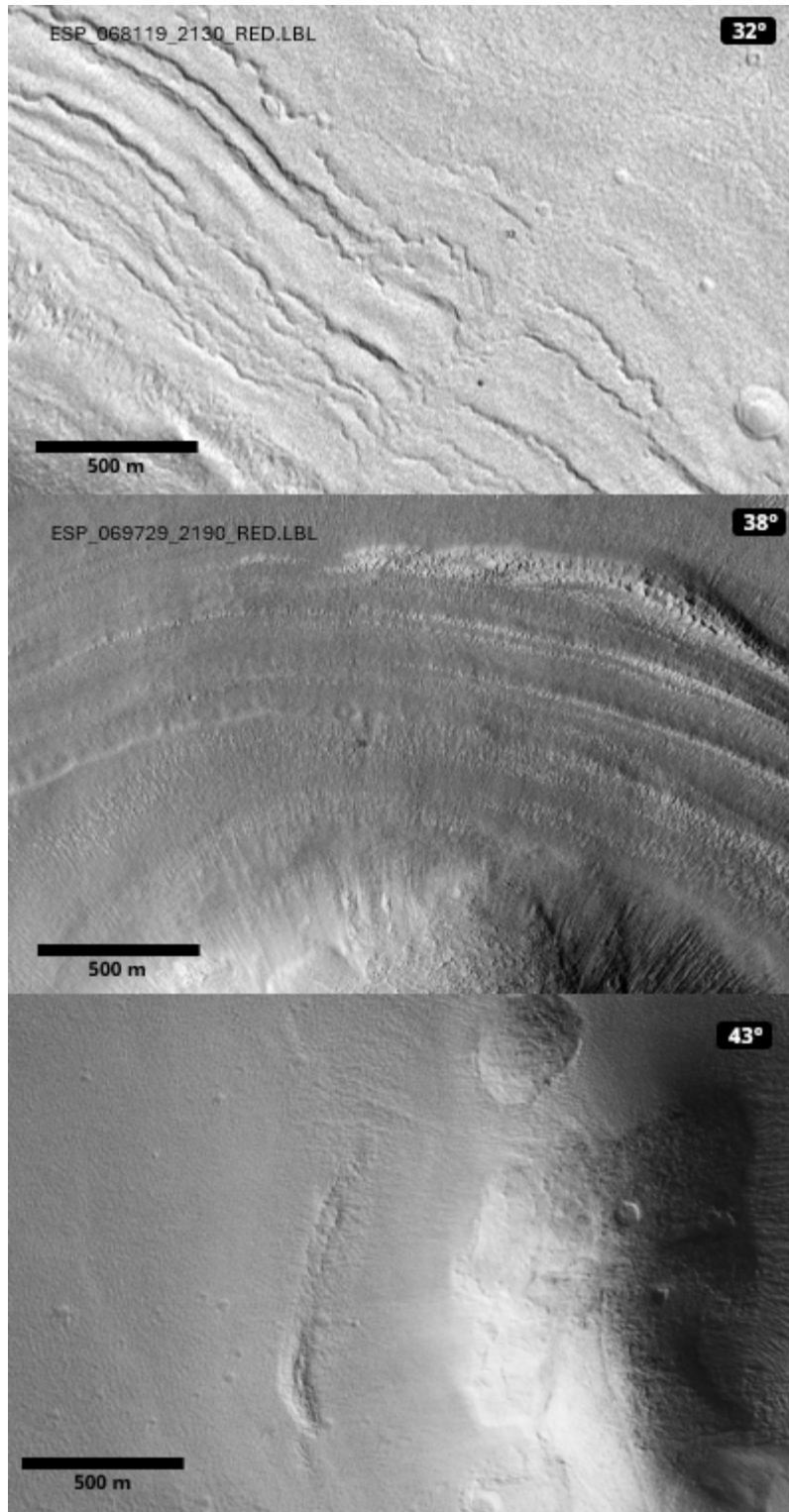


Figure A3. Variations of *Lineated* terrain. Top panel shows a more “deflated” lineated morphology with sharp ridges and edges. Middle panel shows a more concentric, radial pattern with “smoother” ridges. Bottom panel shows an isolated lineation that is hypothesized to have been buried by surrounding terrain. (Source: *HiRISE and/or CTX Global Mosaic v01* if not labeled with *HiRISE ID*).

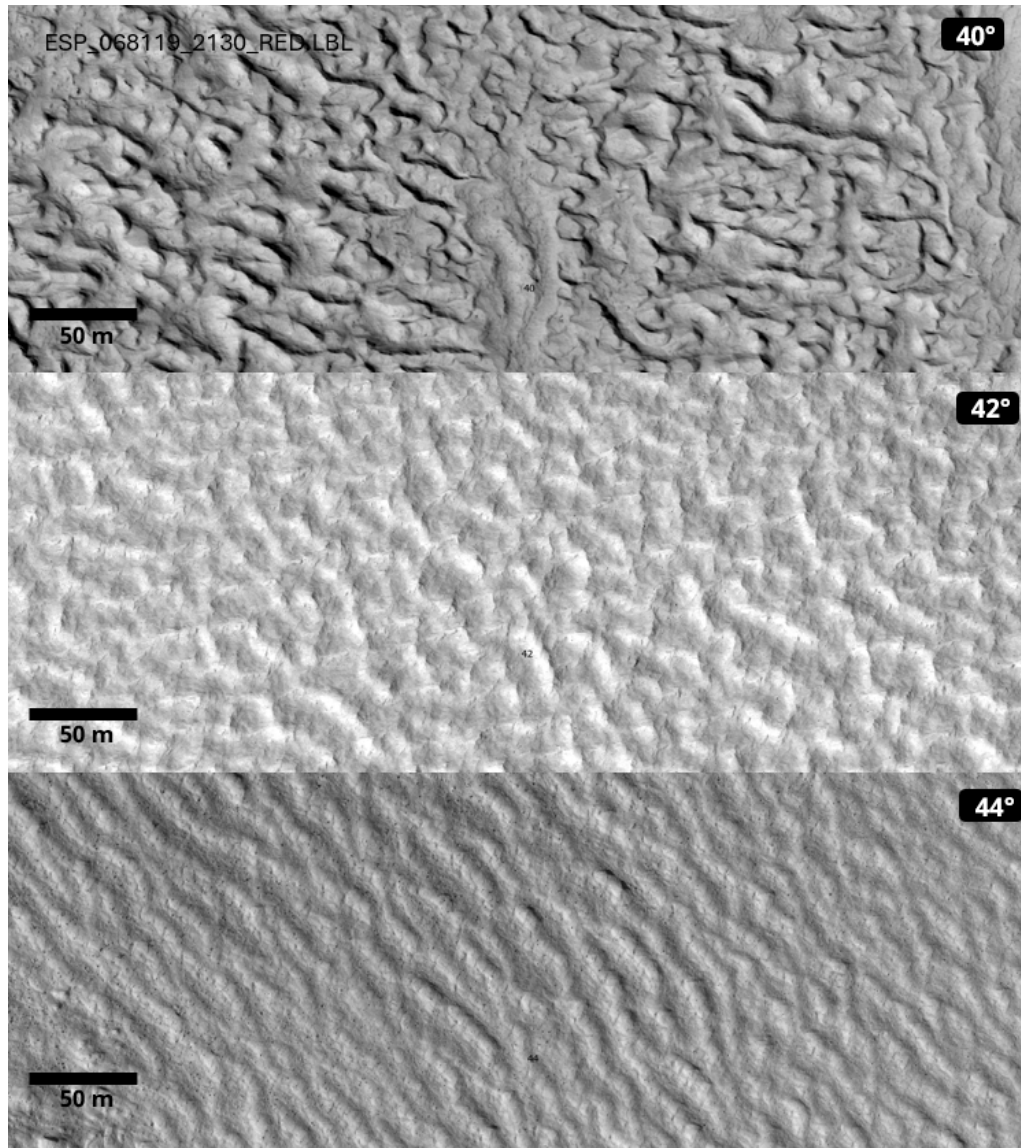


Figure A4. Variations of *Undulating* terrain. From the top to bottom panel, terrain shows degrees of irregularity with the top panel being the most irregular waveform and undulating patterns and the bottom panel being more symmetrical and elongated. (Source: *HiRISE and/or CTX Global Mosaic v01 if not labeled with HiRISE ID*).

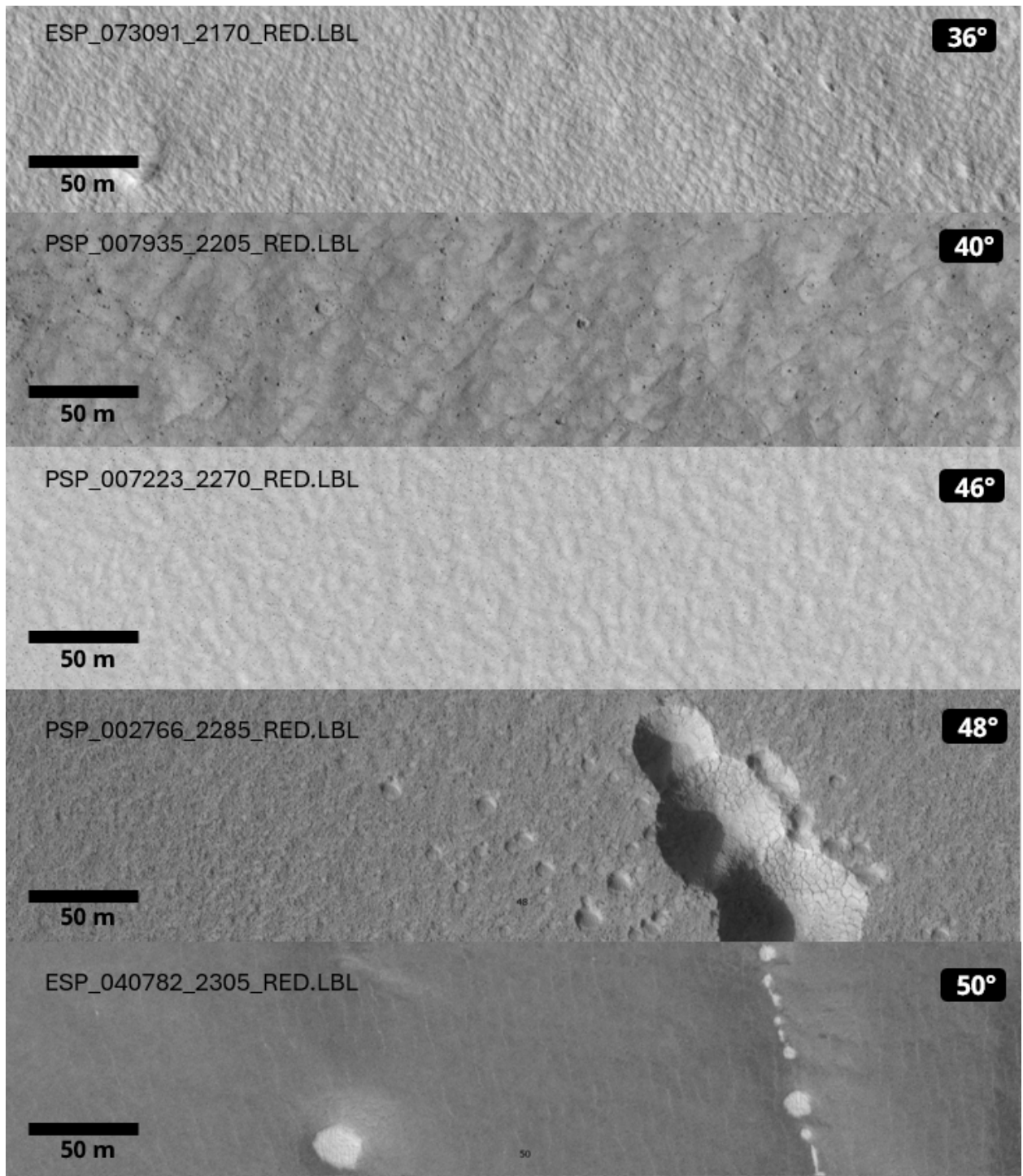


Figure A5. Variations of *Smooth* terrain. All panels show polygonal terrain as a deterministic morphology of *Smooth* terrain with polygons differing in morphology as latitude increases (from top panel to bottom panel). (Source: *HiRISE* and/or *CTX Global Mosaic v01* if not labeled with *HiRISE ID*).

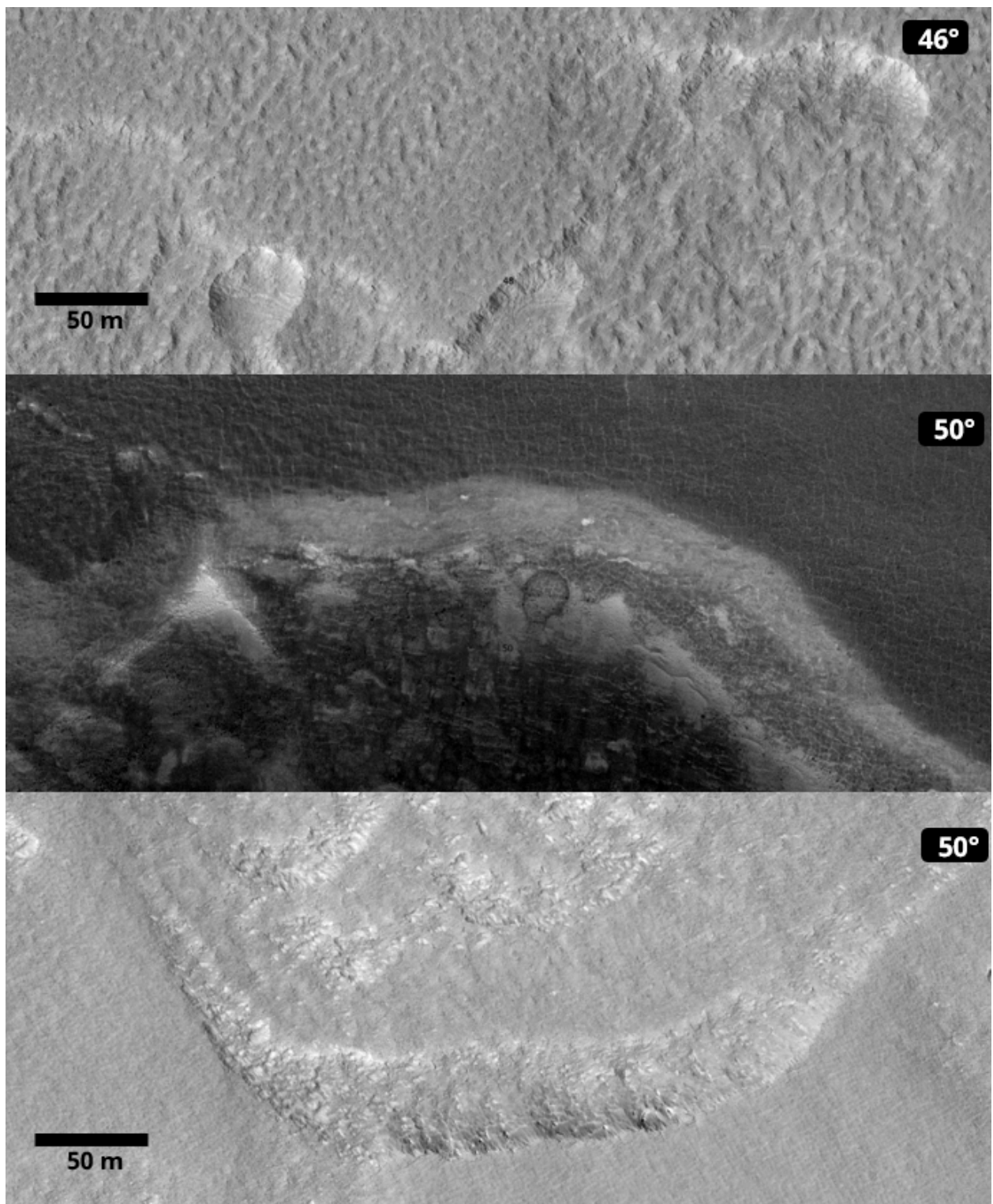


Figure A6. Variations of *Scalloped* terrain. Scallop range from isolated morphologies to coalesced networks of larger scalloped topography. (Source: CTX Global Mosaic v01).

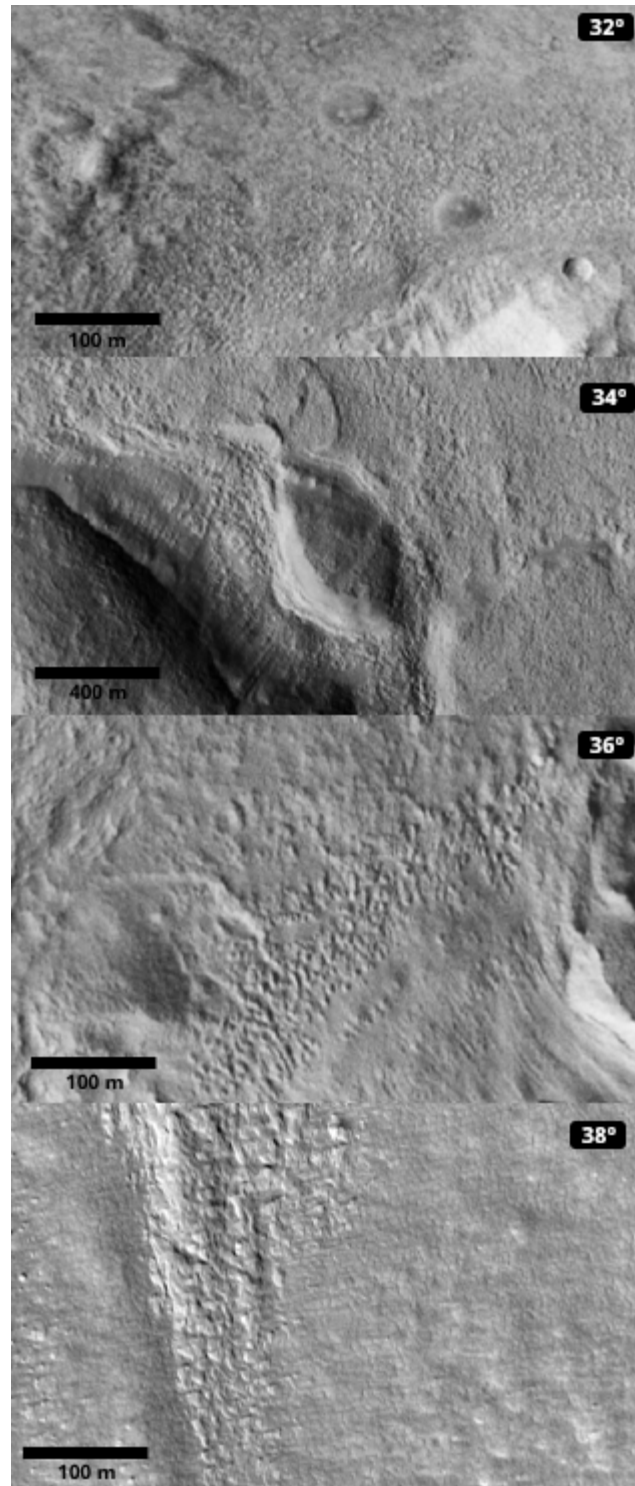


Figure A7. Variations of *Knobby* terrain. This terrain is characterized by its rough, highly irregular surface texture. Unlike other facies, *Knobby* terrain lacks an organized structure and consists of a chaotic assemblage of stippled or mounded landforms. (Source: CTX Global Mosaic v01).