Surface Morphology and Subsurface Ice Content Relationships in Arcadia Planitia, Mars and the Canadian High Arctic

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A thesis submitted in partial fulfillment of the requirements for the Doctor of Philosophy degree in Geology
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Abstract

As NASA and SpaceX prepare for future human missions to Mars as part of an In-situ Resource Utilization (ISRU) Space Act Agreement (SAA), we need more detailed characterization of ice at proposed landing sites to constrain ice accessibility, landing safety, and scientific value. Obtaining near-surface in situ water-ice can be used for rocket fuel and life support needs which would significantly reduce the mass needed for transport to and from Mars. Arcadia Planitia is the lowest-lying region in the northern hemisphere of Mars where abundant evidence exists for an ice-rich subsurface. Shallow Radar observations indicate a decameters-thick layer of water-ice (i.e., buried ice sheet) extends across much of Arcadia. The goal of my Ph.D. research is to characterize the ice-related features at Arcadia Planitia, a proposed future human mission landing site, in detail to assist in the identification of a safe landing site where water-ice is present and accessible for ISRU. By utilizing multiple orbital datasets (i.e., morphology, albedo, thermal infrared reflectance, thermal inertia, and subsurface radar reflections) and identification criteria for Viscous Flow Features (VFFs) on Mars, I mapped six glacial-related features in Arcadia. These units consist of conventional VFFs, such as Lobate Debris Aprons, and non-conventional VFFs. Three sinuous features in the flat-lying plains of Arcadia show surface morphologies and spectral properties indicating these are non-conventional VFFs of channelized ice that once flowed. I propose these sinuous features to be analogous to terrestrial ice streams. Brain terrain is proposed to represent a lag deposit formed atop thick glacial ice as a result of ice sublimation. However, we observe brain terrain to occur only within a narrow latitudinal band within the study site with minimal examples of brain terrain found on the six glacial-related features mapped. We utilize the Canadian High Arctic to investigate analogous brain terrain, that we have termed Vermicular Ridge Features (VRFs), to identify surface-subsurface relationships with ground-penetrating radar, photogrammetry, grain size analysis, and LiDAR. We interpret VRFs to be produced from the passive ablation of stagnant glacial ice. We interpret the lack of brain terrain on the six glacial-related features we mapped at Arcadia Planitia to represent regions where thick units of ice persist, have experienced less degradation than the surrounding terrain, and, therefore, where massive ice is shallower
from the surface making our mapped regions areas where ice is more accessible for ISRU.

**Keywords**

The following keywords can be used to describe this dissertation entitled "Surface Morphology and Subsurface Ice Content Relationships in Arcadia Planitia, Mars and the Canadian High Arctic," which studies the relationship between surface morphology and subsurface ice of glacial and periglacial landforms on Earth and Mars: Geomorphology, Buried Ice, Glacial Geology, Permafrost, Ablation, ISRU, Human Exploration, Mars Polar.

**Summary for Lay Audience**

As NASA and SpaceX prepare to send humans to Mars in the near future, we need to continue to study regions on Mars that are suitable to land. The landing site will need to be safe for landing, scientifically interesting to continue to learn about Mars, and provide significant mineable water-ice as it can be used for life-support and fuel. One area of interest on Mars, called Arcadia Planitia, has abundant evidence that suggests it is both safe for landing and has a large amount of buried water ice. It has been suggested that an ice sheet was buried in the area which is scientifically valuable as glaciers act as physical records of past climate – something we are still trying to understand about Mars. My Ph.D. research aims to describe the surface of ice-related features at Arcadia Planitia in detail and infer areas of minable ice just below the surface to identify an ice-rich and safe landing site for future Martian astronauts to land. This is done by using multiple datasets from Mars that provide information about the surface and subsurface to best locate shallow ice deposits. I mapped six units that I interpret to be buried glaciers on Mars. One unit includes three unique sinuous features in a flat-lying area that I suggest are analogous to ice streams found on Earth. Small-scale ice-related features were also mapped, one of which is called brain terrain and commonly found in the mid to polar
latitudes of Mars. Brain terrain is proposed to represent glacial ice sublimation. I studied a similar landform in the Canadian High Arctic, termed Vermicular Ridge Features (VRFs), to understand the formation process and its relationship with buried ice. We interpret VRFs to form similarly to brain terrain. We found that brain terrain largely does not occur on the mapped glacial units, particularly on the sinuous features. We suggest brain terrain to represent areas where the greatest ice loss has occurred. Therefore, the sinuous features represent areas where less ice loss has occurred suggesting ice may be shallower, easily extractable, and an ideal area to land.
Co-Authorship Statement (where applicable)

Chapter 2: Shannon Hibbard carried out the data collection, georeferencing, remote sensing analysis, mapping, and interpretations of this research, in addition to writing the manuscript. Dr. Nathan Williams assisted in the data collection, georeferencing, remote sensing analysis, and interpretations of this research and provided edits to the manuscript. Dr. Nathan Williams first identified the sinuous features and was the first to suggest they were of glacial origin, while Shannon Hibbard led interpretations for the mechanisms of flow. Dr. Matthew Golombek provided access to Tom Logan’s Mars_Nest software, guidance in remote sensing analysis and interpretations, and edits to the manuscript. Dr. Gordon Osinski provided guidance in remote sensing analysis, data interpretation, and thorough edits to the manuscript. Dr. Etienne Godin provided edits to the manuscript.


Chapter 3: Shannon Hibbard led and decided the field and lab methods approach for this project. Shannon Hibbard processed all data in the lab including orthomosaic, DEM, and GPR processing and interpretation, DEM calibration, satellite data acquisition, and grain size sieving. Shannon Hibbard carried out an extensive literature review, led interpretations, and wrote the manuscript. Dr. Gordon Osinski provided funding for fieldwork, field instrumentation, sample return, software licenses, and summer research. Dr. Gordon Osinski found the landform discussed in this chapter during our 2018 field season. He assisted in field data collection, provided guidance on interpretations, and provided edits for the manuscript. Dr. Etienne Godin assisted in field data collection, provided guidance on interpretations, and provided edits for the manuscript.

A version of this chapter/appendix has been submitted for publication in Geomorphology (Hibbard, S.M., Osinski, G.R. and Godin, E., In Review. Vermicular Ridge Features on Dundas Harbour, Devon Island, Nunavut. Geomorphology).
Chapter 4: Shannon Hibbard was the first to identify the landform on Axel Heiberg Island during our 2019 field season. Shannon Hibbard led and decided the field and lab methods approach for this project. Shannon Hibbard processed all data in the lab including orthomosaic, DEM, and GPR processing and interpretation, satellite data acquisition, and mapping. Shannon Hibbard carried out an extensive literature review, led interpretations, and wrote the manuscript. Dr. Gordon Osinski provided funding for fieldwork, field instrumentation, sample return, software licenses, and summer research. Dr. Gordon Osinski assisted in field data collection, provided guidance on interpretations, and provided edits for the manuscript. Dr. Etienne Godin, Dr. Mark Jellinek, Dr. Antero Kukko, Dr. Shawn Chartrand, Dr. Anna Grau Galofre, and Chimira Andres assisted in field data collection. Dr. Antero Kukko collected and provided LiDAR data for this project.

Chapter 4 is undergoing preparation for submission for publication.

Appendix 1: Shannon Hibbard carried out the data collection, georeferencing, remote sensing analysis, mapping, and interpretations of this research, in addition to writing the report. Dr. Nathan Williams assisted in the data collection, georeferencing, remote sensing analysis, and interpretations of this research. Dr. Nathan Williams proposed a map of small-scale surface morphologies be made. Dr. Matthew Golombek provided access to Tom Logan’s Mars_Nest software, guidance in remote sensing analysis and interpretations. Dr. Gordon Osinski provided guidance in remote sensing analysis and edits to the report. Dr. Etienne Godin provided guidance in the remote sensing analysis.

Appendix 1 is currently formatted as a short report that will be written up into a full paper in the future.
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Graduate school is not easy, but the support from incredible people along the way makes all the difference.

To Dr. Gordon Osinski: I don’t know where to start. Thank you for everything! You have always believed in me, even when I was struggling during my Ph.D. Your supportive, imaginative, encouraging, and professional approach to mentoring has greatly progressed my abilities as a scientist. I have always felt that you have had an undoubted confidence in me in our research which has given me confidence and trust in myself and my abilities. You have provided me with numerous unique opportunities that I am so grateful for and privileged to have taken part in during my Ph.D. This unexpected dash to the finish line would not have been possible without your support, time, and confidence. Thank you so much for everything, Oz. I look forward to continuing to collaborate and do cool science together during my postdoc and beyond.

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<th>Abbreviation</th>
<th>Full Form</th>
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<tbody>
<tr>
<td>Af</td>
<td>Alluvial fans</td>
</tr>
<tr>
<td>AFIDS</td>
<td>Automated Fusion of Image Data System</td>
</tr>
<tr>
<td>Airbus DS</td>
<td>Airbus Defense and Space</td>
</tr>
<tr>
<td>ASTM</td>
<td>American Society for Testing Materials</td>
</tr>
<tr>
<td>Av</td>
<td>Amazonian Volcanic unit</td>
</tr>
<tr>
<td>B</td>
<td>Context Camera primary and extended mission phase beta representing the second Mars year of imaging</td>
</tr>
<tr>
<td>BFR</td>
<td>Big Falcon Rocket</td>
</tr>
<tr>
<td>BP</td>
<td>Before Present</td>
</tr>
<tr>
<td>CCF</td>
<td>Concentric Crater Fill</td>
</tr>
<tr>
<td>CMF</td>
<td>Circular moraine features</td>
</tr>
<tr>
<td>CNES</td>
<td>Centre national d'études spatiales - National Centre for Space Studies</td>
</tr>
<tr>
<td>CO₂</td>
<td>Carbon dioxide</td>
</tr>
<tr>
<td>CTX</td>
<td>Context Camera</td>
</tr>
<tr>
<td>D</td>
<td>Context Camera extended mission phase following mission phase gamma</td>
</tr>
<tr>
<td>DCI</td>
<td>Dust Cover Index</td>
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<tr>
<td>DEM</td>
<td>Digital Elevation Model</td>
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<tr>
<td>DIC</td>
<td>Devon Ice Cap</td>
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<tr>
<td>ESP</td>
<td>Extended Science Phase of HiRISE</td>
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<tr>
<td>Esri</td>
<td>Environmental Systems Research Institute</td>
</tr>
<tr>
<td>Abbreviation</td>
<td>Description</td>
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</tr>
<tr>
<td>fft</td>
<td>fast fourier transform</td>
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<td>g</td>
<td>acceleration of gravity</td>
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<tr>
<td>G</td>
<td>Context Camera extended mission phase gamma representing the third Mars year of imaging</td>
</tr>
<tr>
<td>Ga</td>
<td>giga-annum</td>
</tr>
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<td>GAT</td>
<td>Geospatial Analysis Toolbox</td>
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<tr>
<td>GIS</td>
<td>Geographic Information System</td>
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<tr>
<td>GPR</td>
<td>Ground Penetrating Radar</td>
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<tr>
<td>GPS</td>
<td>Global Positioning System</td>
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<tr>
<td>GRS</td>
<td>Gamma Ray Spectrometer</td>
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<tr>
<td>H</td>
<td>ice thickness</td>
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<tr>
<td>H₂O</td>
<td>water</td>
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<tr>
<td>HiRISE</td>
<td>High Resolution Imaging Science Experiment</td>
</tr>
<tr>
<td>HRSC</td>
<td>High Resolution Stereo Camera</td>
</tr>
<tr>
<td>IDW</td>
<td>Inverse Distance Weighting</td>
</tr>
<tr>
<td>Ih</td>
<td>hexagonal ice</td>
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<tr>
<td>IR</td>
<td>infrared</td>
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<td>ISRU</td>
<td>in situ resource utilization</td>
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<td>J</td>
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<td>JPL</td>
<td>Jet Propulsion Laboratory</td>
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<td>K</td>
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<td>ka</td>
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<td>kHz</td>
<td>kilohertz</td>
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</table>
LAS  LASer file format of lidar point cloud data
LDA  Lobate Debris Apron
LDM  Latitude Dependant Mantle
LiDAR  Light Detection and Ranging
LVF  Lineated Valley Fill
MEPAG  Mars Exploration Program Analysis Group
MEX  Mars Express
MGS  Mars Global Surveyor
MHz  megahertz
MOLA  Mars Orbiter Laser Altimeter
Mr  beach sediments
MRO  Mars Reconnaissance Orbiter
M-WIP  Mining Water Ice on Mars Planning
N/A  Not Applicable
NASA  National Aeronautics and Space Administration
NIR  Near-infrared
NS  Neutron Spectrometer
NSIDC  National Snow and Ice Data Center
OMEGA  Visible and Infrared Mineralogical Mapping Spectrometer
P  Context Camera primary mission phase representing the first Mars year of imaging
p  Power exponent
PSP  Primary Science Phase of HiRISE
Rc  exposed rocky surface
Rr  non-scoured rock
Rs  ice scoured rock
s   seconds
S1  Sinuous feature 1
S2  Sinuous feature 2
S3  Sinuous feature 3
SEC2 Spherical Exponential Calibrated Compensation
SHARAD Shallow Radar
SWIM Subsurface Water Ice Mapping
t  time
T1  terrace 1
T2  terrace 2
T3  terrace 3
TES  Thermal Emission Spectrometer
THEMIS Thermal Emission Imaging System
tiu thermal inertia units
Tmp end moraines with buried glacial ice and ice wedge polygons
Tv  till vaneer
UAV Uncrewed Aerial Vehicle
USDA United States Department of Agriculture
USGS United States Geological Survey
UTM  Universal Transverse Mercator
VFFs  Viscous Flow Features
VRFs  Vermicular Ridge Features
WGS  World Geodetic System

α  ice surface slope
ρ  density of ice
τ  shear stress
τ_d  driving stress

V  Roman numeral 5

Φ  phi units - a logarithmic scale converting mm to whole integers

A  constant largely dependant on ice temperature, crystal orientation, and debris content

n  constant describing the behavior of a viscous material

Ê  strain rate
Chapter 1: Introduction and Background

1.1. Motivation

Compelling evidence for the presence of near-surface water-ice across the mid-latitudes (30–60°) of Mars (e.g., Head et al., 2003; Morgan et al., 2021) yields major implications for the planet’s climate history, total water budget and, consequently, past life and potential for human exploration and habitation. Mid-latitude ice has been of particular interest in the Mars community because it: (1) indicates a change in past orbital, axial, and climate conditions (as ice is not typically stable at the surface in the mid-latitudes), (2) suggests the preservation of ice deposits over millions of years, and (3) is readily accessible as an in-situ resource for human missions based on typical engineering landing constraints (Golombek et al., 2012, 2003).

A high priority goal for the study of Mars outlined in the 2013‒2022 Planetary Science Decadal Survey (Board and Council, 2012) and the Mars Exploration Program Analysis Group (MEPAG) 2020 Science Goals (MEPAG, 2020) is to understand the processes and history of the Martian climate, which is intrinsically related to the deposition, distribution, transportation, and preservation of water-ice. Another high priority goal outlined by MEPAG is to prepare for human exploration by characterizing potentially extractable water-ice deposits for in-situ resource utilization (ISRU).

In 2016, NASA outlined a Mining Water Ice on Mars Planning (M-WIP) study suggesting that the use of in situ water-ice could significantly reduce the total overall mass of fuel needed for transport to and from Mars (AAbbud-Madrid et al., 2016). This study identified important areas of research that will significantly assist in the planning of a future ISRU mission, including to (1) Identify the local characteristics, dimensions, and lateral/vertical continuity of the potential water-ice deposits, (2) Analyze how effectively water-ice deposits can be inferred from a geologic, and possibly Earth analogue, perspective in order to model lateral/vertical variations of the deposit, and (3) Assess the potential for ice equatorward of 50°.
This dissertation encompasses three major themes – namely studies of Martian mid-latitude ice distribution, periglacial and glacial processes on Earth, and remote sensing – to ultimately identify relationships between surface morphology and subsurface ice that can be used to identify where ice exists across the mid-latitudes of Mars and to identify a potential water-ice rich landing site for future ISRU missions.

1.1.1. Goals and Objectives

The goal of this Ph.D. dissertation is to characterize ice in the subsurface of Arcadia Planitia, Mars to assess ice accessibility for ISRU and gain additional insight into Mars’ past climate. This is addressed through the following objectives: (1) Characterize ice-rich landforms remnant from previous glaciations at Arcadia Planitia, (2) Reconstruct the glaciological history and present conditions, including the location and properties of ground ice at Arcadia Planitia, (3) Characterize and propose a formation mechanism for a newly identified landform in the Canadian High Arctic, (4) Utilize this landform as a brain terrain analogue to identify surface-subsurface relationships to assess subsurface ice content and accessibility at Arcadia Planitia.

1.2. Water ISRU

As we continue to push the boundaries of space travel by sending humans to the Moon, Mars, and beyond to not only conduct research, but to establish a permanent presence, we must consider the limitations of current propulsion and life support technologies. Current technology for the propulsion of spacecraft is dependent on, and limited to, the combustion of chemical propellants. A large fraction of the total mass of a rocket is attributed solely to propellants (typically 80–95%), most of which is used just to reach Earth’s orbit from the surface of Earth. The remaining 5–20% of rocket mass includes the rocket’s structure, the leftover fuel, and the payload (typically around 1–2% of the mass), which includes the crew, food and water, research equipment, construction materials, and other necessary equipment. The “tyranny of the rocket equation” (Petit, 2012; Tsiolkovsky, 1965) states that when additional mass is added to the rocket, including fuel, then an even larger amount of additional fuel is required. One method of
compensating for this “tyranny” is by sending multiple rockets that have payloads
dedicated to transporting resources needed for human survival to the proposed
destination, which can be expensive and risky for astronauts relying on successful and
timely resupply missions.

However, utilizing resources outside of Earth can offset the use and cost of resupply
missions, as well as enable commercial business expansion, and facilitate science and
exploration. Water-ice ISRU can be used to produce the propellants needed for fuel
production as hydrogen can act as a reactant and oxygen can as an oxidizer. Liquid
hydrogen and oxygen can also be sold to commercial companies for profit (e.g., Sowers
and Dreyer, 2019). Additionally, accessing resources on another planet will be necessary
for a sustainable future in space exploration.

SpaceX is planning on sending humans to Mars as soon as 2024, with cargo missions to
depart as soon as 2022 (Musk, 2018). The Starship (formerly, BFR) has a large payload
(~10–20% of the rocket mass), which will be possible by refueling in Earth’s orbit to get
to Mars, and then refueling again via ISRU on Mars to return back to Earth (Fig. 1.1).
With an upcoming mission relying on ISRU, it is important that research is focused on
determining the location and attributes of water resource deposits on Mars.

Figure 1.1. SpaceX’s Starship (i.e., BFR) architecture. (1) Ship and tankers are
transported to Earth’s orbit using a reusable booster that returns to Earth. (2) Ship enters
Earth’s orbit. (3) Reusable tankers refill the ship’s fuel in Earth’s orbit until the ship is
full and return to Earth. (4) Refilled ship travels to Mars. (5) Ship is refueled using in situ resources (i.e., water-ice and CO₂) on Mars. (6) Ship launches off of Mars and travels back to Earth. From Musk (2018).

1.2.1. Methods of extraction

Water-ice deposits on Mars have the potential to provide significant water resources for local and long-term extraction (MEPAG, 2020). Multiple methods of the extraction of water-ice on Mars for ISRU have been proposed (Abbud-Madrid et al., 2016) with two options for extracting ice being highly emphasized. An open pit mining method requires removing debris to expose the underlying ice deposit for access to mine (Abbud-Madrid et al., 2016), much like open-pit mining observed on Earth. Shallow ice deposits (2–6 m) are preferred as this method would require significant processing of materials (Abbud-Madrid et al., 2016). Hence, shallower ice would result in less material to process. A down-hole water recovery system, also known as thermal mining, has been proposed which involves drilling into the ice deposit and using a heat probe to melt and/or sublimate the ice (Abbud-Madrid et al., 2016; Sowers and Dreyer, 2019). The recovery system would capture the sublimated water vapor and store it as recondensed ice ready for transport. Both methods rely on confining the thickness of the overburden and knowing the mechanical properties of the overburden and ore deposit. However, orbital datasets on Mars are unable to resolve the upper 10 m of the subsurface, and therefore the use of glacial and periglacial context has been proposed to be an asset in the evaluation of overburden thickness (Bramson et al., 2021; Grau Galofre et al., 2021).

1.2.2. Landing Constraints

Landing in a region where significant accessible water resources are present is critical for the success of future missions targeted for inhabiting Mars (MEPAG, 2020). However, the engineering constraints limit most regions where subsurface and surface ice exist on Mars. Additionally, the scientific value of the landing site must be considered.
1.2.2.1. Safety and Operations

Latitude influences the solar zenith angle and insolation which are important for solar power and providing a manageable thermal environment for equipment to operate. However, insolation directly affects the stability and distribution of water-ice (Madeleine et al., 2009). Consequently, higher insolation accommodates surface operations, but reduces the preservation potential of water-ice and, therefore, reduces ice abundance and complicates ice accessibility. In addition, latitude also affects the fuel efficiency of spacecraft launches. Therefore, the landing site must be located at latitudes ≤ 40° with a preference of being closer to the equator to increase solar capabilities (Fig. 1.2) (Golombek et al., 2021).

Figure 1.2. Map of Mars with major landmarks and regions labeled on a Mars Orbiter Laser Altimeter (MOLA) digital elevation model. From Souness et al. (2012).

Lower elevations provide more atmosphere to slow a descending spacecraft, particularly those with large payloads. With Starship estimated to have a payload greater than Saturn V (~150 tonnes), an elevation below -2 km with respect to the Mars Orbiter Laser
Altimeter (MOLA) geoid is needed, with a preference of elevations below -3 km (Fig. 1.2) (Golombek et al., 2021).

Additionally, low surface slopes and rock abundance, and a radar-reflective and load-bearing surface (i.e., low fine-grained dust content) are necessary to prevent tipping or crashing of the spacecraft during descent and landing (Golombek et al., 2021, 2012, 2003). Therefore, the lowest-lying (i.e., latitude and elevation) region on Mars where significant evidence exists for an ice-rich subsurface is critical for a safe and successful ISRU mission.

1.2.2.2. Scientific Value

Although ice can be used as a resource, it acts as a physical record of the planet’s climate history. Some unanswered questions regarding ice deposits on Mars include, (1) What is the chemistry of the ice and what role does that play in ice stability? (2) What is the relationship between the observed geomorphic features and terrains and the amount and structure of subsurface ice? (3) What emplacement mechanisms and timeframes lead to massive ice vs. pore-filling ice deposits? (4) What is the role of these deposits in creating habitable environments? (Bramson et al., 2021). Therefore, the landing site should ideally include a diverse geologic record to assist in improving our understanding of Mars’ climate record.

1.3. Mid-Latitude Ice on Mars

Changes in the orbital parameters (i.e., eccentricity, obliquity, and precession) of a planet play an important role on climate as they control the distribution of insolation and consequently ice deposition, distribution, and stability. Ice on Mars becomes stable at the surface and near-surface in equatorial and mid-latitudes during high obliquity periods (>30°) and stable at the poles during low obliquity periods (<30°) (e.g., Head et al., 2003; Laskar et al., 2004). Mars’ obliquity has ranged between 15° and 48° over the past 10 Myr and experienced an average obliquity of ~37° over the past 4 Gyr (Laskar et al., 2004), which may have led to multiple cycles of deposition and degradation of mid-latitude ice (e.g., Madeleine et al., 2009). Mars’ obliquity has remained moderately low
(22–26°) over the past 300 kyr, which should result in the degradation of mid-latitude ice (Head et al., 2003). However, evidence of abundant subsurface mid-latitude ice has been observed poleward of 30° by the Gamma Ray Spectrometer (GRS) Neutron Spectrometer (NS) aboard NASA’s 2001 Mars Odyssey (Boynton et al., 2002; Feldman et al., 2002; Pathare et al., 2018), the Shallow Radar (SHARAD) aboard the Mars Reconnaissance Orbiter (MRO) (Bramson et al., 2017, 2015; Morgan et al., 2021), in situ by the Mars Phoenix lander (Mellon et al., 2009), and by a number of morphologic observations of periglacial and glacial landforms made with MRO’s Context Camera (CTX; 6 m/pixel) and High-Resolution Imaging Science Experiment (HiRISE; 0.25–0.50 m/pixel) camera (Levy et al., 2009a; Levy et al., 2014). Therefore, the resulting distribution and concentration of mid-latitude ice are physical records of Mars’ past climate and ice stability in the present climate. By studying the present condition of mid-latitude ice, we can improve our understanding of past and present climate and surface-atmosphere interactions on Mars. This includes determining how much ice is in the subsurface, how deep this ice is from the ground surface, and if liquid water ever played a role in the formation or transportation of ice and ice-related features (Bramson et al., 2021; Grau Galofre et al., 2021).

Recent efforts carried out by the Subsurface Water Ice Mapping (SWIM) project have focused on constraining the distribution of mid-latitude ice on Mars to improve our understanding of the current stability of ice and the global water budget (Morgan et al., 2021). By integrating all relevant orbital datasets, excluding HiRISE, for the investigation of ice, the SWIM project has made significant progress in mapping the distribution of non-polar water-ice in Mars’ northern (Fig. 1.3) and southern (Putzig et al. In Press) hemispheres, indicating exciting new evidence of significant ice deposits in low-lying plains. However, SHARAD measurements are unable to resolve the upper ~5–10 m of the subsurface and other instruments used to detect subsurface ice, such as the GRS, cannot resolve deeper than 1 m. Thus, it is unclear at what depth pure ice is present 1–10 m below the surface. Depth to the top of these shallow ice deposits is extremely important when estimating ice volumes and assessing resource accessibility (Abbud-Madrid et al., 2016).
A variety of glacial (Fig. 1.4) and periglacial landforms can be found across the mid-latitudes of Mars. Viscous flow features (VFFs) (Fig. 1.4), including lobate debris aprons (LDA; Fig. 1.4a), lineated valley fill (LVF; Fig. 1.4b) and concentric crater fill (CCF; Fig. 1.4c), are interpreted to be debris-covered glaciers based on morphology (e.g., Head et al., 2010, 2006a, 2006b, 2005; Souness and Hubbard, 2012) and Shallow Radar (SHARAD) subsurface reflection detections (e.g., Holt et al., 2008; Plaut et al., 2009). Periglacial features, such as polygonally patterned ground, can be found across the mid-latitudes (e.g., Levy et al., 2009a). Other features, such as brain terrain (e.g., Levy et al., 2009b), scalloped depressions (e.g., Dundas et al., 2015), expanded secondary craters (e.g., Viola et al., 2015), and terraced craters (e.g., Bramson et al., 2015) are also commonly found across the mid-latitudes of Mars and have been suggested to indicate the presence of sub-surface ice. However, the amount of ice associated with the formation and preservation of periglacial and glacial features are currently debated.
As we approach the limits of datasets currently available for Mars, observations from Earth analogues become increasingly important. Many comparable ice-related features exist on Earth. By studying periglacial and glacial processes on Earth, we can learn about the formation, preservation and degradation of ice that can help us resolve the upper 10 m of the subsurface on Mars. Ultimately, this will assist in the search for an appropriate landing site for future ISRU and will provide context into Mars’ past climate.

Figure 1.4. Distribution of viscous flow features (VFFs) on Mars. (a) Lobate debris aprons (LDA) as seen in a composite of CTX images P15_007017_2551, P16_007373_2248, and P13_006160_2252. (b) Lineated valley fill (LVF) as seen in a of THEMIS-VIS image V09834018 and Mars Orbiter Camera images E0102224, M0303672, and M0401029. (c) Concentric crater fill (CCF) as seen in a portion of CTX image P15_007028_2164. (d) Distribution of LVF (green), LDA (yellow), and CCF (pink) with an exaggerated apparent area on a 10° latitude/longitude grid over a MOLA topographic base map. From Levy et al. (2014).
1.4. Ice Processes

Much of our understanding of ice processes comes from observations made on Earth, as well as theoretical and lab experiments. Mars has a thin CO$_2$ atmosphere with an atmospheric pressure much lower than that of Earth and experiences very low temperatures that decrease with increasing latitude. These conditions produce a cold and hyper-arid climate not conducive to liquid water, which is an important driver in many ice (i.e., glacial and periglacial) processes on Earth. Nonetheless, it is useful to gain an understanding of ice processes inherent on Earth as the present pressure and temperature conditions on both Earth and Mars support predominantly hexagonal Ice (I$_h$) (Fig. 1.5) and Mars’ past climate has been suggested to have once been more similar to modern Earth (Pollack et al., 1987) or at least provided opportunity for liquid water to exist (Hecht, 2002; Laskar et al., 2004; Marchant and Head, 2007). Additionally, cold and arid climate conditions exist in the polar deserts of the High Arctic and Antarctica where ice processes that are not largely influenced by liquid water can be observed.

![Figure 1.5. H$_2$O and CO$_2$ phase diagram for Earth and Mars. Solid black boxes indicate current average surface temperatures and near-surface atmospheric pressures for Earth](image-url)
and Mars. Larger outlined boxes represent current annual ranges experienced on Earth and Mars based on temporal, elevational and latitudinal variations. Mars ranges are based on Hecht (2002) where liquid water may become present for short periods of time locally. H$_2$O on Earth and Mars are within the pressure and temperature range for the low-pressure crystalline phase of hexagonal ice. From Marchant and Head (2007).

1.4.1. Glacial Processes

Glaciers are perennial masses of land ice that originate from compressed snow which flow when the forces exerted by its weight and gravity on a surface slope overcome the strength of the glacier (Marshall, 2012). Glaciers are often classified based on size, location, and topographic controls, as are glaciers on Mars (e.g., Head et al., 2010). For example, an ice sheet is a mass of glacial ice that extends across more than 50,000 km$^2$ of land, whereas an ice cap covers below 50,000 km$^2$ of land (NSIDC, 2021). A valley glacier is a glacier confined by valley walls, typically in a mountainous region, whereas an ice stream is a glacier confined by slower-moving ice in an ice sheet (NSIDC, 2021).

Glaciers are also commonly classified by their temperature variation, as temperature plays a large role in how a glacier flows, loses mass, and alters the landscape. Pressure increases with depth in a glacier, which, along with impurities, can lower the melting point of ice and is referred to as the pressure melting point. A thermal classification for glaciers is based on the pressure melting point, and includes (1) temperate glaciers, where almost all ice is at or above the pressure melting point (i.e., temperate ice), (2) cold glaciers, where all ice is below the pressure melting point (i.e., cold ice) and frozen to the underlying bed, and (3) polythermal glaciers, which contain both temperate and cold ice.

Based on this thermal classification, the temperature at the base of a glacier is generally classified as being cold-based or warm-based which can be significantly softer and wetter compared to higher in the ice column of a glacier (Hubbard et al., 2003). Basal ice temperatures generally have the most influence on how a glacier flows and modifies the landscape. Glacier motion is caused by ice deformation and basal motion (i.e., sliding over its bed). Internal deformation is driven by shear stress, which, if high enough, will
cause ice to flow plastically and deform under its own weight. A flow law commonly used to describe ice deformation in response to stress acting on a glacier is Glen’s Flow Law (Glen, 1955),

\[ \dot{\varepsilon} = A \tau^n \]

where \( \dot{\varepsilon} \) is the strain rate (i.e., rate of deformation), \( A \) is a constant largely dependent on ice temperature, crystal orientation, and debris content, \( \tau \) is the shear stress (controlled by the density of ice, acceleration of gravity, ice thickness and surface slope), and \( n \) is a constant describing the behavior of a viscous material (usually 3 for glaciers which are a non-linear viscous material).

Shear stress can increase with increasing ice thickness or increasing surface slope. Thereby, the smaller the slope, the more mass is needed to create the same stress as a glacier with a high surface slope. Hence, valley glaciers generally have high surface slopes which provides enough driving stress for their forward motion. However, ice streams often have low surface slopes as they occur in ice sheets, but can experience high driving stresses, such as the Jakobshavn ice stream in the Greenland Ice Sheet (Cuffey and Paterson, 2010). This is due, in large part, to subglacial bedrock topography influencing the driving stress. Yet, ice streams on the Siple Coast of Antarctic experience low driving stresses and low surface slopes, but still generate some of the fastest-flowing ice on Earth (Cuffey and Paterson, 2010; Rignot et al., 2011). This is due to the contribution of basal slip in the forward motion of the ice (Tulaczyk et al., 2000; Winsborrow et al., 2010).

Although basal sliding has been observed at cold-based glaciers (e.g., Cuffey et al., 1999), it mostly occurs in warm-based glaciers. Pressurized water at the base of a glacier reduces frictional forces, essentially reducing bed roughness and shear stress which increases sliding velocity, and therefore glacier velocities (e.g., Iken, 1981; Iken and Bindschadler, 1986). However, if subglacial drainage pathways are well-connected and drainage is efficient, then high water pressures are limited to channels and most of the bed is unaffected by basal sliding (Copland et al., 2003; Willis, 1995).
Subglacial drainage and basal temperatures are important as they affect a glaciers’ erosive capability, and therefore, the landforms left behind. Wet-based glaciers can erode and transport large volumes of sediment which can result in a large variety of landforms, such as moraines, drumlins, scoured bedrock lineated bedforms. Cold-based glaciers experience little movement at the bed, and therefore landforms are typically subdued and more often reflect deposition via surface melt or sublimation (Atkins et al., 2002; Hambrey and Fitzsimons, 2010). Polythermal glaciers would represent an intermediate type producing landforms associated with either or both thermal regimes.

It is currently debated whether VFFs have always been cold-based glacial features (e.g., Head et al., 2010) or if they were once polythermal (e.g., Grau Galofre et al., 2020b) or perhaps even temperate (e.g., Banks et al., 2008; Butcher et al., 2017; Gallagher et al., 2021) in the past. The type of glacier and its subglacial environment holds important implications for Mars’ past climate and its surviving relic landforms. For example, Butcher et al. (2017) and Gallagher and Balme (2015) recently proposed the existence of wet-based glaciation on Mars on the basis of inferred subglacial (waterborne) eskers. However, Grau Galofre et al. (2020a, 2020b) suggested, on the basis of Mars’ low gravity, that such eskers could be formed by localized subglacial meltwater channels rather than by widespread basal sliding (i.e., efficient drainage systems bounded by a frozen ice-bed interface). Evidence is emerging that supports the occurrence of extensive basal sliding and/or subglacial sediment deformation in the presence of water (i.e., wet-based glaciation) on Mars. There is ongoing discussion about the degree to which subglacial hydrology (and spatial variations in the ice-bed coupling it mediates) has influenced VFFs on Mars which would have important implications for Mars past climate and potential for habitability.

1.4.2. Periglacial Processes

Periglacial refers to the conditions, processes and landforms associated with terrestrial materials below 0°C (Washburn, 1980) that is largely dominated by frost action (the process of freeze and thaw of moisture in terrestrial materials) (Harris et al., 1988). Periglacial environments are cold environments that usually contain permafrost (soil,
regolith or bedrock that is at or below 0°C for at least two consecutive years) (Harris et al., 1988) and can be found in the High Arctic polar deserts, tundra zones, northern boreal forests, alpine regions, Antarctic, and high elevation environments of central Asia (Brown et al., 1997). Although moisture in the form of water or ground ice (all types of ice formed in freezing and frozen ground) is a common occurrence in permafrost and drives frost action, it is not required as permafrost is defined on the basis of temperature (Harris et al., 1988).

Two important quantitative parameters of ground ice include the ice content, which refers to the amount of ice contained in permafrost (i.e., the ratio of ice mass to the mass of the dry sample), and excess ice, which is the volume of ice in the ground that exceeds the total pore volume that the ground would have under natural unfrozen conditions (Harris et al., 1988). This is not to be confused with massive ice, which is a comprehensive term used to describe large masses of ground ice, including ice wedges, pingo ice, buried ice, and large ice lenses (Harris et al., 1988). Excess ice includes massive ice; however, massive ice typically has an ice content of at least 250% (Harris et al., 1988) which can be a helpful distinction when classifying ground ice, particularly when glacial ice is left behind and subsequently buried.

Although not specific to periglacial environments, patterned ground is a common surface feature produced by periglacial processes and is widespread on both Earth and Mars. Patterned ground refers to any ground surface exhibiting an ordered morphological pattern, which includes sorted or non-sorted features such as circles, nets, polygons, steps, stripes, and solifluction features (Harris et al., 1988). Sorting refers to the process of soil particles becoming sorted by frost action in the active layer (the near-surface material that is subject to seasonal freeze and thaw). Sorting can occur as a result of frost heaving where frost-induced expansion in the active layer lifts and displaces large grains preferentially. This is due to the size and thermal conductivity of large grains affecting the readiness of moisture to freeze and thaw near it. Stone circles are a common landform formed by the process of frost action and sorting which requires liquid water to be present and these have been suggested as a possible formation mechanism for brain terrain on Mars (Noe Dobrea et al., 2007). However, others (e.g., Levy et al., 2009b) have
suggested brain terrain is associated with and suggests subsurface massive glacial ice on Mars.

Polygonally patterned ground is one of the most common periglacial features on Earth and ubiquitous in the mid-latitudes of Mars. It does not necessarily require liquid water to be present to form. As the ground temperatures decrease, resulting tensile stresses cause fractures that occur as systems of polygons with four or more sides (Washburn, 1980). These cracks can fill with snow, water, and/or sediment while open. When temperatures decrease, the cracks reopen to relieve stresses in the ground and refill. Over time, this can create wedges of ice or sediment in the subsurface and troughs on the surface can grow in width to reflect the existence of a wedge. The growth rate of these wedges is difficult to constrain as it can be influenced by many factors, such as temperature gradient, substrate, moisture content, snow cover and vegetation (Mackay, 1986; Ulrich et al., 2011).

However, a study in the western Arctic in Illisarvik, Northwest Territories, Canada, found that thermal contraction cracks developed at the bottom of a deliberately drained lake just 4 months after lake drainage (Mackay, 1986). This demonstrates that periglacial processes, such as thermal contraction cracking, can begin soon after exposure to subaerial conditions.

There is ongoing discussion about the type of polygons present on Mars (Levy et al., 2010; Mellon, 1997; Soare et al., 2014), which has major implications for past climate and near-surface ice. Polygonal terrain has the potential to host a substantial amount of near-surface ice if thermal contraction cracks are filled with ice or if polygonised soil overlies laterally extensive massive ice, as was observed at the Phoenix landing site (Mellon et al., 2009). However, polygons can appear superficially similar at the surface regardless of subsurface ice content.

1.5. Arcadia Planitia

The lowest occurrence of water-ice detection in the northern hemisphere from GRS (Fig. 1.6) (Pathare et al., 2018) and SWIM results (Fig. 1.3) (Morgan et al., 2021) is found in Arcadia Planitia. Additionally, there is abundant morphologic and radar evidence for
widespread near-surface ice in Arcadia. Evidence of LDAs have been identified in western Arcadia (Mangold, 2003; Plaut et al., 2009). Orbital imaging has revealed recent small impacts that have exposed subsurface ice at mid and high latitudes (Byrne et al., 2009; Dundas et al., 2014, 2021) and erosional scarps exposing massive or excess ice near Milankovic Crater in northern Arcadia (Dundas et al., 2018, 2021). Thermokarstic expansion of secondary craters (i.e., thermal erosion of an existing crater causing the edges to recede) found in the area indicates ground ice sublimation and are referred to as expanded secondaries (Viola et al., 2015). Concentric terracing within the walls of simple craters indicates contrasts in material strengths within the target material where weaker materials are interpreted as ice (Bramson et al., 2015). Polygonal surface patterns, indicative of ground ice and cryoturbation, are also widespread in Arcadia (Williams et al., 2017b).

SHARAD observations made by Plaut et al. (2009), and later expanded on by others (Bramson et al., 2019, 2015; Campbell and Morgan, 2018; Putzig et al., 2019; Morgan et al., 2021), indicate a radar-transparent layer tens of meters thick, that has been interpreted to possibly represent water-ice based on dielectric constants and loss tangents, across much of Arcadia Planitia starting as low as 38°N. Three sinuous features exist where these radar detections suggest thick deposits of ice (Bramson et al., 2015), which are presently interpreted to be relict fluvial channels (Ramsdale et al., 2019).
Figure 1.6. Distribution of water equivalent hydrogen on Mars based on Mars Odyssey Neutron Spectrometer measurements. This map presents the distribution of water equivalent hydrogen (weight %) within the upper meter assuming a two-layer model of a thin layer of regolith overlying a layer of regolith with excess ice. Yellow arrow points to Arcadia Planitia. From Pathare et al. (2018).

The terrain surrounding the sinuous features is characterized by three distinct morphologies, typically found across the mid-latitudes, including polygonally patterned ground (Fig. 1.7a), polygonally patterned brain terrain (Fig. 1.7b), also referred to as brain coral (Noe Dobrea et al., 2007) or crenulated terrain (Williams et al., 2017a), and ground with pitted troughs (Fig. 1.7c). Williams et al. (2017b) reports a potential latitudinal relationship in these features that they suggest is directly related to ice content and stability based on surface morphologies identified in HiRISE images (Fig. 1.7d). The authors suggest that crenulated terrain occurs within a narrow latitudinal band at the southern edge of the study site where the most significant ice sublimation should occur. Polygonal terrain is ubiquitous and is suggested to indicate lower sublimation rates. Polygonal terrain with pitted troughs, referred to as pitted terrain, is found in the northern
latitudes of the study site and indicate localized sublimation. In addition, Williams et al. (2017a) reports crenulated terrain as a version of brain terrain identified on mid-latitude ground rather than overlying only VFFs (Levy et al., 2009b). Bina and Osinski (2021) have reported similar instances of features akin to brain terrain in Utopia Planitia occurring on mid-latitude ground and suggest they are an indicator of ground ice.

Arcadia Planitia is the lowest lying region in the northern hemisphere where significant evidence exists for an ice-rich subsurface and is, therefore, the focus of this dissertation. However, the distinct surface morphologies in Arcadia include polygonal and brain terrain, which are currently debated on their formation mechanism and association with ice. In order to determine if Arcadia Planitia is a potentially favourable landing site for ISRU, the glacial and periglacial history of the region represents a critical research topic to assess ice abundance and accessibility, as well as past climate and scientific value.
Figure 1.7. Dominant surface morphologies in Arcadia Planitia. (a) Polygonal terrain as shown in HiRISE image ESP_037655_2210. (b) Crenulated (i.e., brain) terrain as shown in HiRISE image ESP_036574_2190. (c) Pitted terrain as shown in HiRISE image ESP_028148_2235. (d) Map of color-coded HiRISE images surveyed for dominant surface morphologies overlying THEMIS daytime IR mosaic. Pitted terrain always occurred with crenulated or polygonal terrain. From Williams et al. (2017b).

1.6. Dissertation Outline

As the search for water-ice on Mars for ISRU gathers pace, we must determine a landing site that meets engineering constraints, provides scientific value, and has ice that can be harvested. It is essential to use the current datasets on Mars to their full potential and to
gain insight from Earth analogues where appropriate to best characterize the near-surface of proposed human landing sites. The SWIM project has done exceptional work to improve our understanding of mid-latitude ice distribution on a global scale (Morgan et al., 2021) but each landing site needs to be analyzed at a much higher resolution (e.g., Fig 1.7). This dissertation will provide detailed glacial (Chapter 2) and periglacial (Appendix 1) maps of Arcadia Planitia to identify spatial and subsurface relationships with ice. We utilize the Canadian High Arctic to investigate analogous glacial and periglacial landforms and their relationship with ice (Chapters 3 and 4). By combining observations made at Arcadia Planitia and insight gained from the Arctic, we will assess the resource accessibility at Arcadia Planitia for future ISRU (Chapter 5).

1.7. References


Chapter 2: Evidence for widespread glaciation in Arcadia Planitia, Mars

2.1. Introduction

Evidence of periglacial (e.g., Banks et al., 2008; Byrne et al., 2009; Levy et al., 2009a, 2009b; Mellon et al., 2009b) and glacial (e.g., Head et al., 2010, 2005; Holt et al., 2008; Levy et al., 2007; Plaut et al., 2009b) activity in the mid-latitudes of Mars has been a topic of investigation since Mariner 9 and Viking data revealed features associated with ground ice within “fretted terrain” (Sharp, 1973; Squyres, 1979, 1978). These features were suggested to be ice-related due to evidence of viscous flow in the form of surface lineations, ridges and furrows, and convex profiles (Squyres, 1978) and were later termed viscous flow features (VFFs) (Milliken et al., 2003). VFFs, including lobate debris aprons (LDAs), lineated valley fill (LVF), and concentric crater fill (CCF), have been associated with the presence of ground ice since they were first identified along the dichotomy boundary (e.g., Carr and Schaber, 1977; Squyres, 1979, 1978). The amount of ice in LDA and LVF has been a topic of debate for decades. Early work initially suggested these landforms were composed predominantly of pore-filling ice in erosional debris (e.g., Carr and Schaber, 1977; Sharp, 1973; Squyres, 1979, 1978), with later interpretations as ice-cored masses of debris (e.g., Kochel and Peake, 1984; Pierce and Crown, 2003). However, the availability of subsurface information from the Shallow Radar (SHARAD) sounding experiment on the Mars Reconnaissance Orbiter (MRO) (Seu et al., 2007) has revealed the existence of subsurface ice and these features are now interpreted as large massive ice deposits covered by a thin layer of debris, most analogous to terrestrial debris-covered glaciers (e.g., Head et al., 2006a, 2006b; Holt et al., 2008).

Several criteria have been proposed to identify VFFs (e.g., Head et al., 2010; Milliken, 2003; Souness et al., 2012). These criteria include identifying regions where snow and ice are most likely to accumulate and flow, such as within valleys, alcoves, and around massifs or crater walls and other areas with relief. Features such as lineated, arcuate or compressional ridges of debris are interpreted to be the result of the downslope flow of ice. Previously documented LVF examples are constrained to valleys and can create
valley glacial systems (Head et al., 2010). Thus, conventional VFFs are controlled by topographic constraints and are commonly found at the dichotomy boundary where scarps, valleys and fretted terrain are most abundant (Head et al., 2010; Milliken, 2003; Souness et al., 2012).

In this paper, we report on features reminiscent of LDA and LVF in the plains of Arcadia Planitia, which we refer to as lobate and sinuous features, respectively, and interpret to be glacial in origin (Fig. 2.1). Lobate features occur in eastern Erebus Montes of Arcadia Planitia and are morphologically recognizable as conventional LDA suggesting glacial activity is present in the region (Head et al., 2010). Sinuous features occur in the flat-lying plains of Arcadia Planitia. Ramsdale et al. (2019) suggest the sinuous features described in this study to be relict fluvial channels. However, evidence for massive bodies of extant ice and viscous flow features within the sinuous features leads us to suggest a glacial origin. We created a map of potential VFFs within Arcadia Planitia (Fig. 2.2) by following a broad set of criteria based on those outlined by Head et al. (2010) and Souness et al. (2012). The mapping of more conventional VFFs reminiscent of LDA in eastern Erebus Montes assists in the mapping and comparison to non-conventional VFFs reminiscent of LVF. The sinuous features do not display all attributes of LVF, which we suggest is due to the flat lying nature of Arcadia Planitia where topographic constraints are minimal. The complexity of this area, particularly regarding the sinuous features, lead to several questions: 1) Are the sinuous features LVF? 2) How could snow and/or ice accumulate and flow in flat topography? 3) Are the sinuous features the same age as the local lobate features? 4) What is the relationship between the lobate and sinuous features? 5) Why is eastern Erebus Montes thermally different than the sinuous features? This work is motivated by the goal to better understand the geological and glacial history of Arcadia Planitia and feeds forward to potential future landing sites for human exploration requiring in situ water resources on Mars.
Figure 2.1. THEMIS Daytime infrared mosaic (Edwards et al., 2011; Hill and Christensen, 2017) of map area in Arcadia Planitia. (a) Three bright sinuous features (S1, S2, and S3 labeled with red arrows) are oriented NW-SE and terminate at a NE trending thermal boundary at around 38°N. White box around eastern Erebus Montes showing the location of image b. (b) The eastern Erebus Montes mapping area is characterized by topographic peaks surrounded by lobate features and thermally dark terrain.
2.2. Geological Setting of Arcadia Planitia

Arcadia Planitia lies in the lowlands of the northern hemisphere of Mars in an area of transition between Amazonian volcanics and Hesperian lowlands with Amazonian periglacial modification (Tanaka et al., 2014). The study site is located between 37–43°N and 193–204°E (Fig. 2.1), most of which lies within the Amazonian Volcanic unit (Av) composed of lavas sourced from Elysium Mons or an unknown source northwest of Olympus Mons. This unit is overlain by flood lavas of the Late Amazonian volcanic unit (IAv) in the southern portion of the study area. Tanaka et al., (2014) date Eastern Erebus Montes as part of the Hesperian and Noachian transition unit (HNt) composed of Noachian volcanics and Hesperian mass wasted materials. However, others suggest this area to be Hesperian in age with volcanic sources from Olympus Mons (Fuller and Head, 2002).

Near surface ice on Mars is suggested to become stable at mid-latitudes during periods in which the planet’s obliquity is > 30° (Head et al., 2003; Laskar et al., 2004). Widespread atmospheric deposition of ice across the mid-latitudes is indicated by global climate models during high obliquity periods (Forget et al., 2006; Madeleine et al., 2009). Mars’ obliquity is only ~25° today, and one of the lowest latitude regions where significant widespread near-surface ice is suggested to exist is ~40°N at ~200°E in the smooth plain of Arcadia Planitia (Boynton et al., 2002; Pathare et al., 2018). The SWIM project found ice may extend to as low as 36°N in Arcadia Planitia. However, ice has also been detected to occur as low as ~35°N in Deuteronilus Mensae (Morgan et al., 2021; Piqueux et al., 2019) and Protonilus Mensae (Piqueux et al., 2019). The Mars Odyssey Gamma Ray Spectrometer’s (GRS) Neutron Spectrometer measurements suggests a hydrogen-bearing subsurface layer consistent with 35 ± 15% ice by weight within the first few centimeters of the surface with perhaps more at greater depth (Boynton et al., 2002; Pathare et al., 2018). Arcadia has abundant evidence in support of widespread near-surface ice and periglacial modification. For example, LDA were identified in western Arcadia at the bases of massifs (Head et al., 2010; Plaut et al., 2009b), recent small
impacts exposed subsurface ice at mid- and high-latitudes – including in Arcadia (Byrne, 2009; Dundas et al., 2014, 2021), thermokarstic expansion of secondary craters indicates ground ice sublimation (Viola et al., 2015), concentric terracing within the walls of simple craters suggests discrete layers of ice within the target material (Bramson et al., 2015), and polygonal surface patterns indicative of cryoturbation and ground ice sublimation (Ramsdale et al., 2019; Williams et al., 2017).

SHARAD observations made by Plaut et al. (2009a) indicate a radar-transparent layer up to 100 m thick composed most likely of ice across much of Arcadia Planitia that coincides with the study area, starting as low at 38°N. This work was later expanded on by Bramson et al. (2015) where terraced crater morphology and radar subsurface data were used to define a subsurface layer of water-ice spanning a 1.2 x 10^6 km^2 area of Arcadia Planitia. The authors derived a dielectric constant of 2.5, which they interpret to represent nearly pure ice (which has a dielectric constant of 3) with an overlying regolith, as opposed to a dense basalt (which has a dielectric constant of 8) or more intermediate mixtures of rock and ice (which have intermediate dielectric constants). The occurrence of terraced craters (Bramson et al., 2015), expanded secondary craters (Viola et al., 2015), and fresh craters exposing ice (Dundas et al., 2014) provide morphological evidence of an ice-rich subsurface layer that supports their interpretation of a dielectric constant corresponding to water-ice.

The Subsurface Water Ice Mapping (SWIM) team (Bramson et al., 2019; Putzig et al., 2019; Morgan et al., 2021) expanded on the findings of Bramson et al. (2015) by considering radar, geomorphology and thermal properties to identify subsurface ice in the mid-latitudes of Mars and found that Arcadia Planitia shows a high consistency for subsurface ice (i.e., consistency of multiple datasets suggesting the presence of ice). In fact, in addition to the well-studied valley glacier systems of LVF and LDA at the dichotomy boundary in Deuteronilus Mensae (Head et al., 2010; Petersen et al., 2018), Arcadia shows a higher consistency for ice than other regions in the northern mid-latitudes. The SWIM study suggests a subsurface layer of massive ice (thick units of mostly pure ice that exceeds pore volume) exists in Arcadia that extends farther south than originally thought.
2.3. Materials and Methods

2.3.1. Surveying of Viscous Flow Features

To survey the distribution of surface features and morphologies, we created basemaps composed of geo-referenced images from the Thermal Emission Imaging System (THEMIS) on board the Mars Odyssey spacecraft (100 m/pixel; v.12.0; Edwards et al., 2011; Hill and Christensen, 2017), MRO Context Camera (CTX) (6 m/pixel; Malin et al., 2007) and High-Resolution Imaging Science Experiment (HiRISE) (0.25‒0.50 cm/pixel; McEwen et al., 2007) (Table 2.1). The THEMIS daytime infrared (IR) mosaic was used as a geographic base for all layers in the creation and interpretation of our final map (Figs. 2.1 and 2.2) because thermal data exhibit easily discernible features in Arcadia that serve as control points, while other global-scale visible and topographic datasets do not. A total of 38 CTX images were selected to create a higher resolution mosaic ranging between 37°N to 46°N and 192°E to 205°E. Mars_Nest, a software implemented within Automated Fusion of Image Data System (AFIDS) (Logan et al., 2018), was employed to automatically co-register CTX images to the THEMIS daytime infrared basemap using a fast fourier transform (fft) to create a geo-referenced CTX mosaic. A total of 72 red-filter HiRISE images were then similarly geo-referenced to the CTX mosaic using Mars_Nest and fine-tuned with hirise_shift, a USGS-developed script (Source: Trent Hare). HiRISE images are necessary to resolve the smaller scale surface morphologies and can reveal morphological relationships once presented in the proper geographic context.

Additional datasets were also examined to analyze material, spectral and morphologic characteristics of distinct terrains and their relationships (Table 2.1). All layers are projected in an equidistant cylindrical coordinate system with a planetary radius of 3,396,190 m, a central meridian at 0°, and a standard parallel of 40°, using ArcMap™ 10.5.1–10.7.1. Thermal inertia units (tiu) are in J m⁻² K⁻¹ s⁻¹/². The SWIM team used a median dielectric constant of 4.08 across Arcadia to detect material containing water-ice (Bramson et al., 2019; Putzig et al., 2019). They used this value to estimate depths to the base of the water-ice-rich material, which is the data we rely on here. Depth is reported in meters and indicates maximum ice thickness.
Table 1.1. List of datasets used in this study to characterize the surface and near subsurface of the mapping area in Arcadia Planitia. Instrument acronyms include Mars Orbiter Camera (MOC), Context Camera (CTX), High Resolution Imaging Experiment (HiRISE), Thermal Emission Imaging System (THEMIS), Thermal Emission Spectrometer (TES), Visible Infrared Mineralogical Mapping Spectrometer (OMEGA), High Resolution Stereo Camera (HRSC), Mars Orbiter Laser Altimeter (MOLA), and Shallow Radar (SHARAD). Spacecraft acronyms include Mars Global Surveyor (MGS), Mars Reconnaissance Orbiter (MRO), and Mars Express (MEX).

<table>
<thead>
<tr>
<th>Dataset</th>
<th>Instrument</th>
<th>Spacecraft</th>
<th>Resolution</th>
<th>References</th>
<th>Figures</th>
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<td>MRO</td>
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<td>2.5, 2.10, A2.5–A2.7</td>
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<td>MRO</td>
<td>0.25–0.50 cm/pixel</td>
<td>McEwen et al., 2007</td>
<td>2.5, 2.10, 2.11, A2.3–A2.7</td>
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<td>Mars Odyssey</td>
<td>100 m/pixel</td>
<td>Edwards et al., 2011; Hill et al., 2014</td>
<td>2.1–2.4, 2.6, 2.8, 2.9, 2.12, A2.1, A2.2</td>
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<td>Mars Odyssey</td>
<td>100 m/pixel</td>
<td>Edwards et al., 2011; Hill and Christensen, 2017</td>
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<td>Mars Odyssey</td>
<td>100 m/pixel</td>
<td>Edwards et al., 2011; Hill and Christensen, 2017</td>
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<td>Mars Odyssey</td>
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<td>Edwards et al., 2011; Hill and Christensen, 2017</td>
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<td>MGS</td>
<td>20 pixels/degree</td>
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<td>Ferric/Dust Ratio</td>
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<td>40 pixels/degree</td>
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2.3.2. Criteria for Identifying Viscous Flow Features in Arcadia Planitia

In order to identify VFFs, we considered previously published criteria for identifying LDA, LVF, and other VFFs (Head et al., 2010; Milliken et al., 2003; Souness et al., 2012). We modified these criteria to consider VFFs in the absence of topography. To
remain consistent in our identification of VFFs while mapping, a feature must have the following broad set of criteria:

(i) Be distinct from its surrounding landscape and exhibit a texture or relative brightness different from adjacent terrains,

(ii) Show evidence of flow via surface lineations, ridges or arcuate surface morphologies,

(iii) Have distinct boundaries, including a start, side, or terminus indicating a compositional or process boundary.

2.4. Results and Observations

Six ice-related units were identified based on morphologic, infrared and presence of subsurface radar observations within the study area of Arcadia Planitia (Fig. 2.2). Each unit is described in detail below. The surrounding terrain comprises all unmapped areas north of the thermophysically distinct boundary (Fig. 2.3) within the study site.
Legend

<table>
<thead>
<tr>
<th>Color</th>
<th>Unit</th>
<th>Description</th>
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</thead>
<tbody>
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<td>Unit 1</td>
<td>Sinuous, thermally bright features with a polygonal, ridged and knobby surface texture and arcuate features</td>
</tr>
<tr>
<td>Brown</td>
<td>Unit 2</td>
<td>Lobate features with subdued polygons and near-linear down-slope ridges</td>
</tr>
<tr>
<td>Blue</td>
<td>Unit 3</td>
<td>Rounded cratered hills with boulders</td>
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<tr>
<td>Green</td>
<td>Unit 4</td>
<td>Lobate, gently sloped aprons with near-concentric deformed ridges</td>
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<tr>
<td>Green</td>
<td>Unit 5</td>
<td>Thermally dark mantling unit with extensive brain terrain and linear ridges</td>
</tr>
<tr>
<td>Orange</td>
<td>Unit 6</td>
<td>A smooth textured and polygonised terrain with pitted troughs and abundant expanded secondary craters</td>
</tr>
</tbody>
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End of individual sinuous features for data collection

Unit Interpretation

<table>
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<td>Buried massive ice deposit with a glacial-related origin akin to lineated valley fill</td>
</tr>
<tr>
<td>Lobate debris aprons</td>
</tr>
<tr>
<td>Massifs/exposed bedrock</td>
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<tr>
<td>Ice-cored lobate debris apron moraines</td>
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<tr>
<td>Ice-rich mantling unit</td>
</tr>
<tr>
<td>Transitional ice-rich mantling unit connecting all ice-rich units to one another</td>
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</table>
Figure 2.2. Map of glacial-related terrain in Arcadia Planitia overlying THEMIS daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Individual sinuous features within unit 1 were subdivided to compare properties of each sinuous feature (Table 2.2). The lowest extent of each sinuous feature used for its individual data collection reported in Table 2.2 is indicated by a black dashed line. Exact area used for data collection of individual sinuous features can be found in Figure A2.1. (a) Entire mapping area within Arcadia Planitia, with unit 1 (purple) being the most spatially extensive. The unmapped area is referred to as the surrounding terrain. White box indicates image b. (b) Eastern Erebus Montes sub-region mapped including units 2–6.

2.4.1. Unit 1: Sinuous Features

Unit 1 is made up of three thermophysically distinct sinuous features (S1, S2, and S3) that lie within the lower mid-latitudes of Arcadia Planitia between 37–43°N and 193–204°E (Fig. 2.2). These features trend northwest-southeast (NW-SE) with an unclear beginning and terminus, and run roughly parallel to one another. S1 ranges from roughly 4–40 km in width and extends more than 400 km in length along its central axis from ~43°N–39.5°N before merging with the other sinuous features. S2 ranges from roughly 4–25 km in width and over 400 km in length from ~43°N–38°N. S3 ranges from 2–10 km in width and approximately 200 km in length from ~40°N–38°N. For more representative statistics of unit 1 properties (Table 2.2), the terminus of unit 1 ends at a thermally distinct region running northwest across 37°N–38°N (Figs. 2.2 and 2.3) as the properties south of this boundary are not representative of the properties used to identify unit 1.

Spectrometer data of the study area in Arcadia Planitia suggest a distinct thermal and dust boundary south of the region of interest (Fig. 2.3). This boundary also represents a distinct change in TES and OMEGA albedo (Fig. 2.3). TES datasets indicate this boundary extends slightly farther north than the THEMIS dataset (Fig. 2.3) as a result of earlier data collection from the Mars Global Surveyor (MGS) mission prior to the 2001, 2003 and 2005 dust storms. Sinuous features were separated into individual features for data collection (Fig. A2.1). This excludes tributaries and the southern portion of the map that connects the three sinuous features together.
Figure 2.3. Infrared observations of Arcadia Planitia. (a) THEMIS Daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Thermal boundary near 38°N indicates base of mapping area (outlined by black dashed line). S1, S2 and S3 labeled in red. (b) THEMIS Nighttime IR mosaic (Edwards et al., 2011; Hill et al., 2014). Sinuous features are bight and thermal boundary (dashed black line) is less distinct in night IR. Eastern Erebus region is thermally dark with very thermally bright spots. (c) TES Dust Cover Index (DCI; Ruff and Christensen, 2002). Mapping area has low DCI. South of
boundary (dashed black line) has high DCI (i.e., less dust). (d) TES Bolometric Albedo (Christensen et al., 2001). Mapping area has high bolometric albedo. South of boundary (dashed black line) has low bolometric albedo. (e) OMEGA global nanophase ferric oxide (dust) spectral parameter map (Ody et al., 2012) overlying THEMIS Day IR mosaic with 0% transparency. Mapping area has high dust cover. South of boundary (dashed black line) has low to moderate dust cover. (f) OMEGA global NIR 1-micrometer albedo map reporting Est. Lambert albedo (Ody et al., 2012). Unit 1 shows a higher Lambert albedo than adjacent terrain. Eastern Erebus has the highest Lambert albedo in the area. South of boundary (dashed black line) has low Lambert albedo.

2.4.1.1. Elevation and Topography

The study area where unit 1 occurs (excluding Erebus Montes) is generally flat lying with minimal variations in elevation and gentle slopes (Table 2.2; Fig. 2.4a). Clustered peaks 200–400 m above the plains occur in eastern Erebus Montes (Table 2.2; Fig. 2.4b). The blended HRSC-MOLA DEM indicates S1, S2, and S3 occur within relative topographic lows at their northern ends and become slightly topographically higher (~20–50 m) southward (Fig. 2.4). S1 ranges as low as 20–30 m ±3 m below adjacent terrain northwest of 41.3°N, 200.1°E, and as high as 30–40 m ±3 m above adjacent terrain southeast of 41.3°N, 200.1°E. S2 and S3 generally range between 15–30 m above or below adjacent terrain, in their southern and northern parts, respectively, following a similar pattern as S1 with lower elevations in the northwest and higher elevations in the southeast. Average elevation for unit 1 (-3938 m) is slightly lower than the mapping area average (-3934 m).

2.4.1.2. Morphologic Description

Unit 1 is morphologically distinct from the surrounding terrain (Fig. 2.5). It is characterized by a polygonal, knobby, flat surface with semi-crescentic ridges and landforms consistent with being expanded secondary craters (Figs. 2.5a, b). Polygons in unit 1 are 10–20 m in diameter. Knobs range from 10–20 m in diameter and generally occur in clusters. Semi-crescentric ridges are ~150 m in length and are longitudinal to the
sinuous features’ long axes. It is notable that these ridges and knobs (Fig. 2.5b) occur only within unit 1 and are absent in the surrounding terrain (Fig. 2.5c).

Figure 2.4. HRSC MOLA blended DEM (Fergason et al., 2018) overlying THEMIS daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Elevation is in meters. (a) Study area is generally flat-lying. Sinuous features become higher in elevation farther Southeast. Noise in data increases in the East. White box indicates image b. (b)
Eastern Erebus Montes is characterized by high peaks and is elevated above the low-lying plains of the mapping area.

Figure 2.5. Surface morphologies of unit 1. (a) CTX zoomed image (G20_026091_2182_XN_38N166W) of S3 (labeled). ST = surrounding terrain. A distinct dark boundary (roughly outlined in dashed white), expanded secondaries and arcuate bands (roughly outlined in dashed yellow) are found on S3 and commonly found on the surface of unit 1. Expanded secondary craters are also found in the surrounding terrain but differ in morphology to those found on S3 (compare white arrows in S3 and ST). (b) HiRISE image (ESP_050011_2210) of the surface of S1. Clusters of knobs (green arrow), expanded secondaries (blue arrow), and semi-crescentric ridges (red arrow) are common surface features of unit 1. Images are stretched and true shadows are not cast to determine the height of knobs and clusters. (c) HiRISE anaglyph (ESP_052134_2210_ESP_052358_2210) of S1 (below dashed white arrow) distinct from surrounding terrain (above dashed white arrow). Expanded secondaries visible in
surrounding terrain (blue arrow). Boundary between unit 1 and surrounding terrain indicated by dashed white arrow. This boundary extends in the direction of the arrow.

Linear and arcuate bands are found exclusively within unit 1 (Figs. 2.5 and 2.6), further differentiating this unit from other units and its surroundings. Linear bands can be identified on S2, S3, and isolated areas that comprise unit 1 (Fig. 2.6a-d). Linear bands extend ~72 km along S2 (Fig. 2.6a) and ~30 km along S3 (Fig. 2.6b) along their long axes. These linear bands are most easily seen in THEMIS day or night IR (e.g., Figs. 2.6a, b) but can also be identified in CTX (e.g., Figs. 2.6c, d). S2 also shows linear bands near 42°N, 192.7°E (Fig. 2.6c). This linear band has the shape of a “Y” that branches at 41.8°N and extends up to 23 km southeast. Shadows indicate a slightly raised topography of the linear band, which makes it easy to identify in CTX images. An isolated body of unit 1 near 40.7°N, 199°E (Fig. 2.6d) also shows linear bands with slightly raised topographic relief. All linear bands, excluding S3, have a slightly raised topography with minimal morphologic difference to the rest of the unit. HiRISE images of S3 linear bands reveal that knobs are much less common on the band, and that small circular pits (<20 m diameter) occur instead.

S1 has large near-parallel arcuate bands with a parabolic shape and a vertex directed southeast (Figs. 2.6e–h). These bands are confined within a 5–12 km wide belt that extends from 40°N 150 km southeast before disappearing completely around 37.7°N. S1 bands cross the thermal boundary at 39.2°N, suggesting the terminus of unit 1 extends past the boundary on the map. North of 39.2°N, the bands are less distinct with faint alternating light and dark bands in THEMIS day and night IR. Bands that appear bright in THEMIS daytime IR appear dark in THEMIS nighttime IR (Fig. 2.6e, f). Arcuate bands can be seen in CTX and HiRISE by alternating light and dark banding (Fig. 2.6g) or slight bulging of the bands (Fig. 2.6h). South of 39.2°N, bands become more easily identifiable with alternating light and dark bands in CTX and in THEMIS day and night IR. The bands become increasingly irregular in morphology farther south (Figs. 2.6e–g). Arcuate bands on S3 are much smaller and morphologically distinct (Fig. 2.5a). They are
confined to a 2.5 km belt that extends 8.2 km to the southeast. S3 bands are difficult to identify in THEMIS day and night IR, but easily distinguishable in CTX by an outline of small ridges (e.g., Fig. 2.5a). HiRISE anaglyphs (ESP_052134_2210_ESP_052358_2210 and ESP_057633_2200_ESP_066521_2200) indicates the bands have slightly raised rims. Rims are roughly 240 m wide. No light and dark banding is present.

2.4.1.3. Infrared Observations

The THEMIS Quantitative Thermal Inertia Global Mosaic reports night infrared thermal inertias in thermal inertia units (tiu; J m$^{-2}$ K$^{-1}$ s$^{-1/2}$). The study area has an average value of 169 tiu, maximum of 595 tiu and a minimum of 25 tiu with THEMIS thermal inertias (Table 2.2, Fig. 2.7a, b). TES daytime-derived thermal inertias (Fig. 2.7c) of the study area are comparable but provide a much lower maximum of 339 tiu. However, TES nighttime-derived thermal inertias (Fig. 2.7d) indicate a maximum of 1158 tiu in the study area which is found in S2. Unit 1 comprises the highest average thermal inertias in THEMIS and TES datasets with S3 having the highest average based on THEMIS and S1 having the highest average based on TES in both day and night (Table 2.2). Unit 1, and its individual sinuous features, exhibit relatively high thermal inertia values in both TES day and night ±20 tiu (Table 2.2). THEMIS daytime and nighttime IR maps give a visual representation of these high values in both day and night (Figs. 2.3a, b, 2.7a).

Individual sinuous features were subdivided in unit 1 to compare thermal inertia statistics (Fig. 2.2 and A2.1; Table 2.2). Major tributaries were removed from the sinuous features and each were terminated before the thermal and dust boundary at ~37° and 38°N (Fig. 2.3) to collect the most representative values of each sinuous feature.

Arcuate bands in S1 alternate between ~216 tiu and ~197 tiu THEMIS-derived thermal inertias. THEMIS Day and Night IR show an inversely proportional relationship between alternating light and dark arcuate bands (Fig. 2.6e, f). Linear banding found in S2 and S3 have a similar relationship where THEMIS IR day and night are inversely proportional. S2 bands are bright in the day and dark at night, whereas S3 bands are dark in the day and bright at night.
Figure 2.6. Longitudinal linear (white arrows) and transverse arcuate (white dashed lines) bands found in unit 1. All figures oriented where north is up. (a) THEMIS Daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017) zoomed on S2. Linear bands indicated by white arrows. Image center at 39.835, 196.035E. (b) THEMIS Daytime IR mosaic zoomed on S3 with median linear band (white arrow). Image center at 38.290, 195.231E. (c) CTX image (F02_036640_2216_XN_41N167W) of northern portion of S2. Y-shaped linear band indicated by white arrow. Image center at 41.865, 198.701E. (d) CTX image (G03_019524_2211_XI_41N161W) of linear band (white arrow) on isolated section of unit 1. Image center at 40.910, 198.962E. (e) THEMIS Daytime IR mosaic zoomed on arcuate bands of S1 below the thermal boundary. General shape of band outlined in white dashed line. Image center at 38.712, 203.893E. (f) THEMIS Nighttime IR mosaic (Edwards et al., 2011; Hill et al., 2014) of arcuate bands in Fig. 2.6e. General shape of band outlined in white dashed line. (g) CTX image (G02_018825_2189_XN_38N156W) of alternating light and dark arcuate bands in Figs. 2.6e and f. General shape of band outlined in white dashed line. Image centered at 38.694, 203.845E. (h) CTX image (G03_019392_2203_XN_40N156W) of northern portion of S1 arcuate bands. General shape of band outlined in white dashed line. Image centered at 39.584, 203.499E.
Figure 2.7. Thermal imaging of mapping area in Arcadia Planitia. (a) THEMIS Daytime IR with Night IR colorized (Edwards et al., 2011 and Hill and Christensen, 2017). Warmer colors (reds and yellows) indicate warmer nighttime temperatures and higher thermal inertias. Cooler colors (blues and greens) indicate cooler nighttime temperatures and lower thermal inertias. (b) THEMIS Thermal Inertia overlying THEMIS Daytime IR with 0% transparency (Christensen et al., 2013; Fergason et al., 2006). Varying saturation values are due to seasonal, atmospheric, and geometric viewing conditions. Overlapping images and global mosaics show inconsistencies in the absolute measured values of thermal inertia. (c) TES Daytime Thermal Inertia (Putzig and Mellon, 2007). (d) TES Nighttime Thermal Inertia (Putzig and Mellon, 2007).
TES albedo (Fig. 2.3d) and dust cover index data (Fig. 2.3c), and Viking rock abundance data, are too coarse to identify differences between the sinuous units and their surrounding terrain. However, TES bolometric albedo indicates high values (0.14–0.25) across Arcadia (Table 2.2), with the exception of the thermally distinct band running slightly northeast just below the sinuous units. TES dust cover index (DCI) shows optically thick dust overlies most of Arcadia except for the thermally distinct band just south of the sinuous units (Fig. 3c). The mapping area has an average of 0.94 DCI. HiRISE images show very few rocks > 1 m in diameter occurring on the sinuous units and the surrounding terrain (e.g., ESP_051923_2210 and ESP_051857_2210).

The OMEGA global NIR 1 μm albedo map (Fig. 2.3f) indicates that unit 1 has a high Lambert albedo with an average of 0.35, maximum of 0.38, and minimum of 0.24 (Fig. 2.3; Table 2.2). Yet, the surrounding terrain has higher Lambert albedo values to unit 1 (Fig. 2.3f; Table 2.2). The lowest Lambert albedo values occur at the base of the mapping area with values around 0.24.

TES datasets indicate the thermal/dust boundary is different than the boundary seen in THEMIS and OMEGA datasets (Fig. 2.3). This is likely due to earlier data collection of the Mars Global Surveyor (MGS) mission carrying TES. THEMIS and OMEGA datasets were collected during later missions on the Mars Odyssey and Mars Express, respectively. Global and local dust storms occurring in 2001, 2003, and 2005 resulted in dust deposition and redistribution of this thermal/dust boundary. However, this relict boundary is identifiable in THEMIS daytime IR (Fig. 2.8).

Table 2.2. Statistics of all datasets used for study site in Arcadia Planitia. Surrounding terrain excludes all mapped units. Map area includes all units and surrounding area. S1, S2, and S3 excludes tributaries. Sinuous features’ southernmost extent is outlined in Figure 2.2. Erebus Montes excludes unit 1 and surrounding terrain. Area is calculated using a Mars Equidistant Cylindrical SP40 projected coordinate system. TES and OMEGA dust values excluded from table due to coarse resolution. However, some values are reported in the text. tiu = thermal inertia units = J m$^{-2}$ K$^{-1}$ s$^{-1/2}$. SHARAD data from
SWIM team indicated base of a water-ice-rich material (Bramson et al., 2019; Putzig et al., 2019).

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### SHARAD

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</tr>
</tbody>
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48
2.4.1.4. Radar Observations

Subsurface reflections from SHARAD (Bramson et al., 2019; Morgan et al., 2021; Putzig et al., 2019; Seu et al., 2007) have been interpreted to possibly be the base/bottom of an ice rich layer. The depth of this layer has been proposed to be up to ~114 m which would put an upper limit on the layer thickness (Table 2.2; Fig. 2.9). Subsurface reflections along S1 appear to deepen southward with shallow depths reaching 15 m and maximum depths reaching up to 72 m. S2 is only partly intersected by radar transects that indicate depths reaching between 15 and 60 m. S3 is intersected by one pixel of the SWIM subsurface radar interface map suggesting a depth of 20 m is reached. The average depth reached in unit 1 is 38 m. In general, subsurface reflections to the base of the ice-rich layer are comparable under the sinuous features and surrounding terrain but S1 has deeper subsurface reflections than the surrounding terrain on average.

2.4.2. Units 2-6: Erebus Montes

Erebus Montes consists of 5 units determined on their morphology, thermal properties, and subsurface radar reflections. Additional morphological context into each unit can be found in Appendix 1 (Fig. A2.2–A2.7).
Figure 2.8. SWIM team map showing estimated depth to base of ice interface in meters overlying THEMIS Daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Unit 1 boundary outlined in white. (a) Deep subsurface reflections occurring within S1 compared to adjacent terrain. (b) Few data points in eastern Erebus region. Data from Bramson et al. (2019), Putzig et al. (2019) and Seu et al. (2007).
2.4.2.1. Elevation and Topography

Eastern Erebus Montes covers an area of 6,666 km² is higher in elevation compared to the rest of the mapping area (Fig. 2.4). Major topographic variations exist (Figs. 2.2b and 2.10a, b) here with the highest elevations occurring in unit 3 with a maximum elevation of -3727 m. Unit 2 is in contact with unit 3 and transitions into unit 4 which then transitions into unit 5 with decreasing elevations. Unit 6 reaches the lowest elevations in Erebus with a maximum of -3666 m, which is approximately 100 m higher than unit 1 maximum elevations (Table 2.2).

2.4.2.2. Morphologic Description

Units 2–4 comprise lobate features that occur between 38–40°N, 193–197°E. Unit 2 is characterized by a relatively smooth surface with subdued polygons ranging between 10 and 14 m in width, features consistent with expanded secondaries, and subtle linear slope-parallel ridges (Fig. 2.10c and A2.3). These slope-parallel ridges always occur at the boundary between units 2 and 3. Other parallel linear forms can sometimes be found at the base of unit 2 (Fig. A2.3a). These linear forms are approximately 2 times wider than the slope-parallel ridges and are not always slope-parallel. Cracks can occur and run parallel to the unit 2 and unit 3 boundary (Fig. A2.3b, c). The boundary between units 2 and 3 are determined by a change in slope, drop in elevation, and/or change in surface texture (Fig. 2.10 and A2.2a–d). Unit 2 is sometimes separated from unit 3 by a depression, commonly referred to as a “moat,” or scarps (Fig. 2.11; A2.4a).

Unit 3 is always topographically higher than its surroundings (Fig. 2.4) and is characterized by a cratered, rounded (Fig. A2.4a) to flat (Fig. 2.11), and blocky surface (Figs. 2.11 and A2.4). Craters in unit 3 do not show signs of expansion. If an outcrop of unit 3 has a scarp face, it will have a gradual transitional boundary on the opposite side (Fig. 2.11). A series of parallel lineated grooves and ridges can be found on 2% (4 outcrops total) of unit 3 (Fig. 2.11 and A2.4d). The azimuth of these features is inconsistent across the 4 individual outcrop examples. These linear features can be aligned NW-SE, NE-SW or directly S and these lineations occur on flatter surfaces only.
The boundary between units 2 and 3 can be clearly identified by a change in topography, change in relative brightness, presence of a moat, change in crater expansion, and/or by unit 2 morphologic indicators described above (Fig. A2.4).

Unit 4 exhibits multiple textures but the presence of semi-concentric sinuous ridges is most diagnostic (Fig. 2.10d and A2.5a). These sinuous ridges create circular raised ridges in places, commonly referred to as deformed ridges and pits by Pierce and Crown (2003) (Fig. A2.5a). Unit 4 also exhibits smaller raised anastomosing ridges and troughs similar to brain terrain (Levy et al., 2009) or crenulated terrain (Williams et al., 2017) (Fig. 2.10d and A2.5b–d). Crater-like features consistent with expanded secondaries can be also be found in unit 4. The boundary of unit 4 is not always clear and can range from sharp and easily distinguishable by a sudden change in morphology and relative brightness (e.g., Fig. 2.10 and A2.5b, d), to very gradual and unclear, particularly when adjacent to unit 5. Unit 4 exhibits parallel linear features when transitioning to or in contact with unit 6 (Fig. A2.5d).

Unit 5 is the most difficult unit to characterize. It is best characterized as a unit that connects individual outcrops of units 2–4 to each other as it shares many features with unit 4. Bumpy terrain (Fig. 2.10e) and parallel linear forms (Fig. A2.6a, b) like those seen on units 2 and 4 are commonly found on unit 5. These linear forms are oriented roughly NW–SE. Crenulated terrain is also common on unit 5 (Fig. A2.6a). The easiest way to identify the boundary between units 4 and 5 is by noting the farthest extent of deformed ridges (e.g., A2.6c). Unit 5 can most clearly be identified in THEMIS day and night IR (Fig. 2.1b, 2.2b, and 2.3b) but can also be identified by a lobate feature or textural change from the surrounding terrain (Fig. A2.6c, d). This boundary is additionally identified using elevation data where a drop in elevation occurs at the outermost extent of unit 5 (Fig. 2.4).

Unit 6 is also a difficult unit to characterize. It most frequently forms a transitional unit between all units (Figs. 2.2), particularly between unit 1 and Erebus (e.g., A2.7a). Unit 6 is characterized by a relatively smooth, polygonized and dark surface with many features consistent with expanded secondaries (Fig. A2.7). Unit 6 can be difficult to distinguish...
from unit 1, but it generally appears smoother with no ridges and can appear relatively brighter than unit 1 in HiRISE and CTX (Fig. A2.7a). The boundary between unit 6 and the surrounding terrain or other units can be distinct and easily identifiable by a change in morphology and relative brightness (Fig. A2.7b, c). It can sometimes appear to overlie units 2–4 (e.g., Fig. 2.10). However, this is rarely observed and can more commonly be found adjacent to unit 4 rather than on top of unit 4 (e.g., Fig. A2.7c). Unit 6 can also more gradually transition into the surrounding terrain with patches of surrounding terrain occurring within unit 6 near the boundary (e.g., Fig. A2.7d). This gradual transition is apparent through a change in morphology and relative brightness.

Figure 2.10. Surface morphologies of map units 2–6. Refer to legend in Figure 2.2. (a) Unmapped CTX image mosaic (P17_007736_2198_XN_39N166W and
D22_035796_2198_XN_39N164W) centered at 39.604, 197.783E in eastern Erebus. (b) Mapped version of Fig. 2.10a. Unit 2 = brown, unit 3 = blue, unit 4 = teal, unit 5 = green, unit 6 = yellow. Legend in Fig. 2.2. (c) HiRISE image (PSP_009740_2200_RED) of units 2 and 3 at zoom. Unit 3 on the top right corner of the image. Unit 2 with slope-parallel ridges and expanded secondaries. Image centered at 39.592, 194.799E. (d) Unit 4 exhibiting semi-concentric sinuous ridges and crenulated/brain terrain. Image centered at 39.578, 194.775E. (e) Unit 5 exhibits bumpy terrain. Image centered at 39.548, 194.751E.

2.4.2.3. Infrared Observations

THEMIS-derived thermal inertias for eastern Erebus (units 2–6) are lower compared to unit 1 with an average of 146 tiu (Table 2.2, Fig. 2.7). Among units in eastern Erebus, unit 6 has the highest average thermal inertia of 172 tiu, followed by units 3 and 2 (Table 2.2). Units 4 and 5 have the lowest average thermal inertias in eastern Erebus but have relatively high maximum values. The highest individual thermal inertia values for eastern Erebus can be found in unit 2 with a maximum of 590 tiu. Unit 2 also has one of the lowest thermal inertia values in Erebus of 27 tiu (Table 2.2).

Unit 6 has the highest average TES daytime (140 tiu) and nighttime (169 tiu) derived thermal inertias in eastern Erebus, followed by units 2 and 3 (Table 2.2). Unit 4 has the lowest average thermal inertias of the Erebus region, which is comparable to values for unit 5.

TES bolometric albedo shows little variation among units in Erebus, with an average of 0.24 in units 2–6. Erebus units have a greater minimum bolometric albedo of 0.23 compared to unit 1 (0.14) and the surrounding terrain (0.16) (Table 2.2). However, OMEGA Lambert albedo indicates eastern Erebus has higher Lambert albedo values than the other units, with an average of 0.37, maximum of 0.52, and minimum of 0.32 (Table 2.2, Fig. 2.3d). Highest and lowest individual Lambert albedo values occur in units 2 and 3 and unit 4 has the highest average Lambert albedo of 0.38 according to OMEGA data.
This difference in values may be due to resolution, wavelength differences in the datasets, or and/or scattering properties. TES bolometric albedo should remove most small-scale effects of slope as thermal wavelength calculations are linear.

TES DCI and OMEGA ferric dust resolution are too coarse to resolve individual units. However, TES DCI in Erebus has an average of 0.93, maximum of 0.97, and minimum of 0.89. OMEGA dust in Erebus has an average of 1.01, maximum of 1.03, and minimum of 1.

Figure 2.11. HiRISE image (ESP_017243_2195_RED) showing surface morphologies found on some massifs of Unit 3. Green arrow indicates smooth transition from unit 2 lobate debris apron into unit 3 massif. Red arrow indicates the moat separating unit 2 lobate debris apron from unit 3 massif. Blue arrow shows scarp on southern end of unit 3 massif. Black dashed line outlines linear grooves on the surface of unit 3 massif. Image centered at 38.998, 195.445E.
2.4.2.4. Radar Observations

Deep radar reflections mapped by the SWIM team do not intersect the majority of eastern Erebus but intersects some of the map units (Fig. 2.9b). SWIM subsurface interpretation map pixels are large with a length of ~3 km and a width that ranges between 2.3 and 7 km. This variation in pixel width is due to the clipping process of the data to determine general statistics for each unit (Table 2.2). ArcGIS attempts to clip the data to fit within the specified map unit which shortens the width of the pixel. This process minimizes pixel overlap; however, overlap still occurs over multiple units in Erebus (Fig. 2.9b). Subsurface reflections indicate the base/bottom of a possible ice-rich layer reaches depths up to 100 m in Erebus (Table 2.2; Fig. 2.9). Units 2 and 3 reflections reach up to 100 m. Units 4 and 6 have equivalent maximum depths of 83 m but different averages (Table 2.2). Unit 4 has deeper subsurface reflections on average than units 5 and 6. Erebus units have comparable reflection depths to S1 (Table 2.2).

2.4.3. Surrounding Terrain

The surrounding terrain includes all unmapped area in the study site (Figs. 2.2 and 2.5). This area consists of polygonal terrain and brain terrain (described by Williams et al., 2017). It lacks any evidence of viscous flow and therefore does not meet the criteria outlined in section 3.2. Thermal inertia varies widely in the surrounding terrain and yields a slightly lower thermal inertia average, maximum and minimum than unit 1 (Table 2.2). The deepest SHARAD returns occur in the surrounding terrain (Table 2.2; Fig. 2.9). The in-depth analysis of surface morphologies and other orbital characteristics observed in the surrounding terrain are outside the scope of this study and are the focus of a separate study (Appendix 1).
2.5. Discussion

2.5.1. Unit 1: Sinuous Features of Arcadia Planitia

2.5.1.1. Debris-Covered Glacier Origin

The simplest explanation for unit 1 is that the sinuous features (S1, S2, and S3) are LVF, as they resemble this VFF and occur in close spatial association with lobate features, later interpreted to be LDA. LVF are a common VFF found in the valleys of fretted topography and are most well known for their longitudinal lineated striations of debris at the surface. LVF are longer than they are wide and are interpreted to be flowing masses of debris-covered glaciers, similar to the valley glaciers of the Antarctic (Head et al., 2010). Many features of unit 1 are shared with conventional glacier-like forms (GLF) described by Souness et al. (2012) and VFF described by Head et al., (2010).

Unit 1 has distinct boundaries, a criterion used to identify GLF (Souness et al, 2012), discernable by TES and OMEGA albedo, relative brightness in HiRISE and CTX, morphology and thermal properties (Figs. 2.5 and 2.7). For example, a distinct morphology found only on unit 1 are clusters of knobs (i.e., Fig. 2.5b), which may be similar to knobs commonly found on LDA (e.g., Pierce and Crown, 2003), referred to as “knobby texture,” or a series of pits and small buttes formed from thermokarstic degradation (Mangold, 2003). However, small mounds can also be produced from porewater expulsion to produce pingos (e.g., Dundas et al., 2008; Dundas and McEwan, 2010; Lefort et al., 2009), the inversion of secondary craters (e.g., Dundas and McEwan, 2010), lava and ice or groundwater interactions to form rootless cones (Lanagan et al., 2001), or hydro- or cryovolcanic eruptions (Gallagher et al., 2018). Nevertheless, evidence for ice in the region is widespread and these knobs are not found outside of the sinuous features, thus we suggest that these knobs are likely most akin to the knobby texture found on LDA or inverted secondary craters.

Additionally, evidence of viscous flow is indicated by transverse arcuate bands found on S1 and S3 and longitudinal linear bands observed on S2 and S3 (Figs. 2.5a and 2.6).
Surface foliation, such as compressional or extensional ridges, surface lineations, and transverse arcuate ridges or lobes, are characteristic features of LDA and LVF (Head et al., 2010; Souness et al., 2012) and have been interpreted as flow-deformed ridges of debris due to glacial flow of a debris-covered glacier (Head et al., 2010). If these arcuate ridges and longitudinal linear bands are indeed flow-deformed ridges of debris, this may explain subtle differences in thermal inertia and thermal reflectance of the linear and arcuate flow bands compared to the rest of the sinuous feature. These flow-like features are not found in the flat-lying regions surrounding the study site. The only other place flow features occur in the study site are on the isolated masses of unit 1, and the LDA of Erebus Montes where topographic variations are much more pronounced. These flow-like features, along with the sinuous nature of unit 1, indicate a general flow direction to the southeast.

Other common features of LVF, such as the tightening and folding of flow-deformed ridges to create complex folds where flow is constrained due to topography or where LVF merge (Head et al., 2010), are not apparent in unit 1. We suggest that the absence of these other common LVF features is likely due to the lack of topographic variation in the area to act as flow barriers or that periglacial modification has destroyed these features.

Further evidence for the glacial origin of the sinuous features comes from subsurface reflections observed by SHARAD from the SWIM team, which imply the presence of massive ice in the region (Table 2.2; Fig. 2.9). Transects suggest deeper subsurface reflections under S1 than under the surrounding terrain and similar ice thickness averages exist for S1 and Erebus where LDA are suggested to occur. A rough outline of a radar transparent layer, interpreted to be subsurface ice, includes the area where we mapped unit 1 (Plaut et al., 2009a). In addition, a radar map produced by Bramson et al. (2015) shows increased depths to the base of a potential ice layer in our mapping area within Arcadia Planitia. This is further supported by the work of Bramson et al. (2017) who suggests an ice-rich layer analogous to a remnant ice sheet is buried in Arcadia Planitia. Finally, widespread expanded secondaries suggest ice in the shallow subsurface of the study area (Bramson et al., 2015; Viola et al., 2015). Unit 1 secondaries are consistently expanded, indicating near surface ice is likely continuous spatially across the unit.
Based on our observations of the thermal behavior of unit 1, the sinuous features are most obvious in thermal infrared, appearing bright in both day and night (Figs. 2.3 and 2.7). Unit 1 has moderate thermal inertia but contains the highest average day and night thermal inertias in the study area (Fig. 2.7, Table 2.2). The TES DCI is relatively low (indicating significant dust) and TES albedo is relatively high in the study area, indicating that Arcadia is blanketed with an optically thick layer of dust across the region and abruptly becomes less dusty at the ~38°N boundary (Figs. 2.3 and Table 2.2). The moderate thermal inertia of unit 1 is too high to represent thermally thick (cm to tens of cm) loose dust and too low to represent pure ice or rock. The low rock abundance, lack of boulders identified in HiRISE images, and moderately low DCI, along with comparisons to surface materials at existing landing sites, argues that the surface of unit 1 is covered by common Martian regolith composed dominantly of sand that is either cohesionless or with low cohesion (Golombek et al., 2008). The remnant of a relict dust boundary on S1 (Fig. 2.8) in addition to the persistence of a near-linear dust-free boundary following multiple dust storms (Fig. 2.3) indicates a small amount of sintered or ice-cemented dust resistant to complete removal from dust storms may unevenly blanket the surface of the sinuous features, or possibly the entire mapping area north of the thermal boundary (Fig. 2.3a, b).

Based on observations of the Phoenix landing site, shallow ground ice contributed to thermal inertias between 190 and 325 tiu with an average of 250 tiu (Mellon et al., 2009a). Seasonal, atmospheric, and geometric viewing conditions also affect measurements of thermal inertia, and overlapping images and global mosaics show inconsistencies in the absolute measured values of thermal inertia. Orbit-based derivations of the thermal inertia of ground ice are also complicated by seasonal variations in frost. The thermal inertia values can be fit by either a low thermal inertia regolith/sand layer over a high thermal inertia ground ice or rock layer, or by a slightly elevated thermal inertia regolith/sand without ice. Other aspects of unit 1 provide stronger support for shallow ground ice. Models of cryoturbation to form polygonal terrain and scaling of observations made at the Phoenix landing site scale to suggest the regolith cover in Arcadia is on the order of 10 cm thick (Mellon et al., 2009a, 2008). Mellon et al. (2008) predicts ice-table depths greater than 20 cm in Mars’ current climate.
conditions would have insufficient thermal contraction to result in polygonal cracking. However, they also suggest an ice-table depth of less than 6 cm would result in 5 m spacing between cracks, smaller than the 10‒20 m polygon diameter observed in Arcadia. Thus, the presence of polygonally patterned ground in Arcadia would suggest a regolith/sand layer thickness between 6 cm and 20 cm.

Assuming an ice-rich deposit exists 6 cm below the surface and extends downward to an average depth of 37.7 m, as suggested by SHARAD data (Table 2.2; Fig. 2.9), unit 1 could consist of ~2,091 km³ of water-ice. However, this is a very rough estimate and should be considered with caution. S1 has the most SWIM interpretation map coverage within unit 1, with an average depth to an ice-rich base of 41.22 m, yielding a rough volume estimate of ~478 km³. The proposed volume of water-ice in Arcadia Planitia is 13,000‒61,000 km³ (Bramson et al., 2017) with unit 1 comprising ~3.5‒16% of the ice.

The source and timing of the deposition of widespread, thick deposits of near-surface ice in Arcadia Planitia is unclear. However, many studies agree that the burial and preservation of past snowfall accumulation may have resulted in massive ice in Arcadia Planitia (Bramson et al., 2015; Viola et al., 2015; Bramson et al., 2017, 2019; Madeleine et al., 2009; Putzig et al., 2019 Ramsdale et al., 2019), which is a typical suggested source for VFFs on Mars (Head et al., 2010).

In summary, we interpret unit 1 to be a debris-covered glacier, much like a conventional LVF. However, the sinuous features of unit 1 have some distinct differences from LVF making them unique from any other VFF seen on Mars. Unit 1 is not confined or controlled by surface topography, differentiating these features from conventional and previously documented LVF or fretted terrain. Therefore, we look to Earth for analogous examples.

2.5.1.2. Ice Stream Origin

The forward motion and flow of ice in a flat landscape has been observed in Antarctica, which is commonly compared to Mars due to its stable cold and dry conditions (e.g., Dickinson and Rosen, 2003; Marchant and Head, 2007). Ice streams are regions in an ice
sheet or ice cap where ice moves more rapidly than its surrounding ice (Stokes and Clark 1999). Two types of ice streams occur: topographic ice streams and pure ice streams (Stokes and Clark, 1999). Topographic ice streams are constrained and largely controlled by subglacial bedrock topography, whereas pure ice streams are not. Ice streams are some of the fastest moving ice in all of Antarctica (Rignot et al., 2011) even if gravitational driving stress is low, such is the case with pure ice streams. The origin of their fast flow is still largely unknown but is commonly attributed to basal pressurized water, saturated basal till, or subsurface topography (Alley et al., 2004; Winsborrow et al., 2010).

Stokes and Clark (1999) outline 8 geomorphological criteria for assisting in the identification of paleo-ice streams on Earth, including (1) characteristic shape and dimensions (high aspect ratio), (2) highly convergent flow patterns, (3) highly attenuated bedforms, (4) Boothia-type erratic dispersal trains, (5) abrupt lateral margins, (6) lateral shear margins, (7) pervasively deformed sediment, and (8) offshore sediment accumulation.

The sinuous features in Arcadia Planitia exhibit at least 2 (possibly 4) of the 8 key indicators, including a characteristic shape and dimension that has a high aspect ratio (i.e., Fig. 2.2), and abrupt lateral margins (i.e., Fig. 2.5), which may represent channelized ice (Fig. 2.12). The sinuous features also show additional key indicators, such as convergent flow patterns (i.e., Fig. 2.6c) and highly attenuated dispersal trains (i.e., Figs. 2.6a–d). They lack key features such as lateral shear margins and surface crevasses. However, these are features that may become buried following stagnation of an ice stream (e.g., Retzlaff and Bentley, 1993). Thus, a possible analogous interpretation of unit 1 is a stagnated and subsequently buried ice stream system possibly overlain by the LDM and dust.

It is important to keep in mind that the characteristics presented by Stokes and Clark (1999) are helpful for identifying geological imprints of paleo-ice streams, but we are suggesting that the sinuous features may be more analogous to stagnant ice streams, and therefore never receded or fully ablated. Therefore, streamlined bedforms, for example, are not expected to be present. Additionally, Stokes and Clark (1999) suggest that
defining characteristics are often modified or destroyed so it is common to only find 2–3 characteristics in a potential paleo-ice stream on Earth, but that size, shape and abrupt margins may be the most telling features.

Although ice streams have not been extensively discussed on Mars (e.g., Lucchitta, 1982, 2001), their ability to reach high ice velocities often with no apparent influence from surface topography, in addition to their characteristic morphology, makes them an interesting mechanism to consider for the sinuous features of Arcadia Planitia. This would have major implications for Mars’ past climate and would provide further evidence that Martian mid-latitudes were extensively glaciated. Global climate models have suggested widespread accumulation of snow in the mid-latitudes at high obliquity (Forget et al., 2006; Madeleine et al., 2009), which is consistent with ice sheet formation. Fastook et al. (2014) suggests the collapse of mid-latitude cold-based ice sheets and subsequent covering of debris from exposed bedrock is what formed and preserved LDA and LVF.

The burial of a decameters thick ice sheet made of predominantly pure ice sourced from snowfall accumulation is suggested to exist in Arcadia Planitia, including our study site, based on SHARAD radar reflections, dielectric constants, orbital evolution and thermal stability simulations, and vapor migration models (Bramson et al., 2015, 2017). This potential buried ice sheet may be directly related to the proposed ice streams, suggesting that Arcadia Planitia may be the site of a relict buried glacial landsystem. Deposition of the LDM (Head et al., 2003) would have followed and allowed the buried ice to remain stable for millions of years (Fastook et al., 2014; Bramson et al., 2017).
Although ice stream basal flow dynamics are not fully understood, it is suggested that basal water plays a significant role in both the initiation and flow of an ice stream by either sliding over a lubricating layer of water or deformable water-saturated till (Alley et al., 2004; Bell et al., 2007; Winsborrow et al., 2010). However, many factors are hypothesized to control the initiation, location and velocity of ice streams, including subglacial topographic focusing, subglacial topographic steps, macro-scale subglacial topographic roughness, calving margins, subglacial geology, geothermal heat flux, and subglacial meltwater routing (Alley et al., 2004; see review in Winsborrow et al., 2010).

Wet-based glaciation in Mars’ recent past is an ongoing topic of debate (e.g., Fassett et al., 2010; Fastook and Head, 2015; Gallagher and Balme, 2015; Grau Galofre et al., 2020a, 2020b; Head and Marchant, 2003; Levy et al., 2016). Banks et al. (2008) has suggested the action of wet-based glaciation in the southern mid-latitudes of Argyre Planitia is responsible for the presence of features interpreted to be lineated grooves, U-shaped valleys, boulder deposits, cirques, and streamlined hills. Additionally, eskers, a

![Figure 2.12. Earth analogue comparison. (a) Synthetic Aperture Radar (SAR) derived images of Ice Stream B (Whillans Ice Stream) in Siple Coast, Antarctica from Alley et al. (2004). (b) Sinuous feature 2 (S2) in THEMIS Daytime IR (Edwards et al., 2011; Hill and Christensen, 2017) in Arcadia Planitia, Mars.](image-url)
glacial feature that can only be produced from subglacial drainage, have been proposed in the mid-latitudes of Mars (e.g., Butcher et al., 2017; Gallagher and Balme, 2015).

Some suggested sources of subglacial melting on Mars includes local geothermal heat flux anomalies (Gallagher and Balme, 2015), subglacial volcanic eruptions (Scanlon et al., 2015), warmer global climate (Scanlon et al., 2018), and lowering of the melting-point of ice due to impurities (Greve et al., 2004). However, Grau Galofre et al. (2020a, 2020b) suggest glacial sliding is largely subdued on Mars due to its low gravity. The authors suggest low gravitational driving stress would result in subglacial drainage rather than subglacial sliding on Mars. However, a deformable ice-sediment interface can lead to rapid sliding (Winsborrow et al., 2010). Additionally, models by Kleiner and Humbert (2014) suggest that Antarctic ice stream motion is more heavily reliant on ice and bedrock geometry rather than basal water content. They suggest that strain heating from ice deformation at the base of a cold-based glacier can produce enough heat to incorporate microscopic water inclusions into the basal ice. This small inclusion of water creates a warmer, more deformable ice that could still create surface velocities observed in the western Antarctic ice streams without basal sliding. The channelized nature and the indication of viscous flow in the sinuous features identified in Arcadia Planitia may be a result of subsurface topography.

The SWIM team found that deep reflectors they interpret as the base of an ice-rich layer only partially covers the study area (Fig. 2.9). Where these deep reflectors are present, subsurface reflections appear to be deeper underlying southern portions of unit 1 compared to its adjacent surroundings. This indicates a potential increased ice thickness in the sinuous features and could suggest a subsurface topographic control initiating and constraining the flow of ice into a channel-like form. When ice is channelized through a topographic trough, velocity increases to maintain a constant discharge of ice. Additionally, thicker ice generates more frictional heat which either enhances the probability of basal meltwater production or lowers the viscosity of the ice to promote greater ice deformation (Boulton et al., 2003; Clarke et al., 1977; Cuffey and Paterson, 2010; Winsborrow et al., 2010). Moreover, high basal shear stress along an ice stream can effectively minimize the importance of basal sliding and other basal motions to
produce fast flow (Clarke and Echelmeyer, 1996). For example, Jakobshavn Isbrae Ice Stream, a topographic ice stream in western Greenland, largely relies on internal deformation of the ice under large driving stresses to reach high flow velocities (Clarke and Echelmeyer, 1996). However, ice deformation alone does not normally produce ice stream velocities observed on Earth. The combination with basal sliding is suggested to play an important role (Cuffey and Paterson, 2010; Winsborrow et al., 2010).

On Earth, we see a range of driving stresses. For example, Whillans has a driving stress of 15 kPa and Jakobshavn has a driving stress of up to 410 kPa (Cuffey and Paterson, 2010). When driving stresses are low, basal slip caused by pressurized water or weak till beds becomes important for ice stream initiation and velocity (e.g., Tulaczyk et al., 2000b). If a bed is too strong (i.e., basal freeze-on, rough or non-deformable till or bedrock), internal ice deformation and basal shear can still occur, but low driving stresses tend to lead to slow ice velocities not typically observed in terrestrial ice streams (Tulaczyk et al., 2000a). The Siple Coast Ice Streams, which tend to be thinner and wider than topographic ice streams, are suggested to attain their high flow velocities via lateral margin shear and basal slip over an undrained water-saturated weak till (Tulaczyk et al., 2000a). Although unpopular, it has been argued that Whillans Ice Stream largely flows due to internal ice deformation with the basal 3 cm frozen to the ground (Bennett, 2003; Kamb, 2001). This suggests that the fast flow of an ice stream with low surface slopes may not necessarily rely on basal sliding and subglacial water. However, Glen’s power law for glacier ice flow demonstrates an exponential relationship between temperature and ice deformation. As temperature of the ice decreases, the ease at which the ice can deform exponentially decreases, which increases the importance of subglacial sliding for the forward movement of ice. Nevertheless, ice stream flow dynamics are still debated and unclear.

A simple calculation for driving stress, \( \tau_d = \rho g H \sin \alpha \), can be done with some assumptions, where \( \rho \) is the density of ice (assuming 1200 kg m\(^{-3}\) calculated by Zuber et al. (2007) for a dominantly water-ice mass with 15% silicate dust content), \( g \) is the acceleration of gravity (3.711 m s\(^{-2}\)), \( H \) is the ice thickness (using max. thickness of 72 m), and \( \alpha \) is ice surface slope. However, the sinuous features indicate flow is to the SE.
(Fig. 2.6), but elevation increases to the SE (Fig. 2.4a). This suggests the current conditions did not produce the flow features observed and thus the driving force was different at the time these flow features were produced. Thus, we cannot calculate a representative driving stress or suggest a reliable flow mechanism (i.e., topographic vs. sliding) as ice thickness and surface slope were likely different in the past.

In summary, we propose that the sinuous features of unit 1 are channelized ice most analogous to terrestrial ice streams that once flowed (Fig. 2.12). The mechanism of ice motion is unclear but may largely flow via internal ice deformation as is suggested for other VFFs (Fassett et al., 2010). However, with the absence of surface topography contributing to a high driving stress, this necessitates the subglacial topographic focusing of the ice to yield a high enough driving stress that exceeds high basal friction without the presence of subglacial water. Therefore, we should not discount that basal slip or basal deformation of a water-saturated fine-grained layer may have contributed in part to its forward motion. Pure ice streams, which seem to heavily rely on basal slip, are highly variable as they change directions and initiate/stagnate often. Therefore, if the sinuous features did rely on slip, they may have been short-lived and may not have required significant water supply. It is also common for an ice stream to shut down at different points along its length, as suggested to have happened to the Kamb Ice Stream (Retzlaff and Bentley, 1993), which may explain the discontinuity in flow features across unit 1.

Further insight from modeling of the flow of channelized ice under low stress conditions on Mars is necessary as it yields important implications for Mars’ climate. The presence of ice streams on Mars may significantly change our understanding of glacial dynamics, landsystem evolution, and climate.

2.5.1.3. Non-Glacial Origins

Ramsdale et al. (2019) interpreted the sinuous features in our study area to be relict fluvial channels with large quantities of fluvially-sourced ice either by the freezing of wet fluvial sediments or of shallow groundwater. Viola et al. (2015) and Ramsdale et al. (2019) suggest the growth and build up of ice lenses may contribute to some excess (ice
volume that exceeds pore volume) subsurface ice in Arcadia as well. Ramsdale et al. (2019) suggest these channel systems flowed to the northwest where they branched into smaller channels. However, this interpretation does not explain the viscous flow features indicating a flow direction to the southeast. The morphological and thermal properties that comprise unit 1 in this study were not considered by Ramsdale et al. (2019). Based on our mapping of the study area, unit 1 begins as sinuous features that merge farther south, and it contains isolated outcrops in the mapping area, which are not reflected in the annotated map of the channelized terrain to which Ramsdale et al. (2019) refers. To expand on this interpretation, the sinuous features could be interpreted as relict eskers with subglaciofluvial-sourced ice. However, there have been two examples of eskers tied to source VFF features (Butcher et al., 2017; Gallagher and Balme, 2015) neither of which appear to contain ice or have a similar morphology to the sinuous features.

Kargel et al. (1995) identified similar sinuous features farther west in Arcadia Planitia (~44°N, 176°E) in lower resolution (10s to 100s m/pixel) image data and interpreted them as tunnel channels and eskers formed from widespread mid-latitude glaciation. These sinuous features described by Kargel et al. (1995) in higher resolution images most resemble S3 in appearance by their narrow and sinuous nature and the presence of a medial longitudinal flow line. However, the sinuous features described by Kargel et al. (1995) appear to occur within troughs (i.e., CTX image P20_008910_2245_XI_44N184W). Examples of eskers on Mars are increasing (e.g., Banks et al., 2009, 2008; Butcher et al., 2016, 2017; Gallagher and Balme, 2015; Head, 2000; Head and Hallet, 2001; Kargel and Strom, 1992; Scanlon et al., 2018), which are identified as sinuous ridges of sediment deposited from subglacial fluvial systems. Grau Galofre et al. (2020) suggests eskers are more likely to be identified on Mars rather than other wet-based glaciation landforms, such as scour marks and moraines. The authors suggest that the low gravity on Mars would inhibit sliding and thus features indicative of drainage, such as eskers, would be more feasible. However, eskers differ from unit 1 sinuous features in morphology and scale and do not explain the viscous flow features observed throughout unit 1. There is strong evidence to support the claim that unit 1 sinuous features are composed of ice rather than sediment. The proposed eskers identified by Kargel et al. (1995) may be additional ice-related sinuous features rather than eskers.
Higher resolution imagery reveals more morphologic similarities between these and the unit 1 sinuous features described here than with candidate eskers (e.g., Butcher et al., 2017). Their surface morphology appears flat, polygonal and knobby with distinct boundaries (i.e., ESP_054416_2240_RED) that intersect features resembling LDAs (i.e., B17_016136_2245_XI_44N182W).

Channelized lava flows (e.g., Baloga et al., 2003; Garry et al., 2007) and lava tubes (e.g., Léveillé and Datta, 2010) have been identified on Mars and, given the geologic context of Arcadia (Tanaka et al., 2014), we must also consider a volcanic origin for the sinuous features. Arcadia is overlain by flood lavas, although the source is debated (Fuller and Head, 2002; Tanaka et al., 2014). Lava rheology, effusion rate, and slope are factors which control flow behavior and morphology of a lava channel (Dietterich and Cashman, 2014). Lava channel morphology can vary widely, but generally have long and narrow forms, with levees (e.g., Bailey et al., 2006; Garry et al., 2007). Dietterich and Cashman (2014) suggests that longer-lived channels are more likely to be narrow and have higher levees. However, levees can collapse or erode, which could explain the absence of levees on the sinuous features. Lava tubes are confined in the subsurface and create relatively isolated linear channels (e.g., Léveillé and Datta, 2010) that morphologically differ from observations of the sinuous features. Dielectric constants of deep radar returns in unit 1 (Bramson et al., 2015) are too low for there to be a significant rock component and are similar to that of water-ice. The abundance of radar, thermal, and morphologic evidence for an ice-rich subsurface in Arcadia leads us to interpret unit 1 as glacial-related rather than volcanic.

2.5.2. Units 2–4: Lobate Debris Aprons in Erebus Montes

2.5.2.1. Evidence from Multi-Dataset Mapping

Lobate debris aprons have been suggested to occur in Arcadia Planitia (Head et al., 2010; Plaut et al., 2009b), but have not yet been characterized in detail at this location. Pierce and Crown (2003), Head et al. (2010), and Souness et al. (2012) have created a set of criteria for the identification of LDA, LVF, and other conventional VFFs, all of which
suggest flow ridges and lineations are key indicators of glacial influence. Pierce and Crown (2003) describe LDAs as exhibiting geomorphic features and surface textures, such as blocks, alternating slope-parallel grooves and ridges, pits, moats, cracks, deformed rims, ridge-and-valley texture, knobby texture, and sharp-ridge texture. Head et al. (2010) and Souness et al. (2012) make note of other important features in LDA, such as flow-deformed ridges and flow lineations as a product of compressional and extensional forces from the flow of viscous material downhill. As we outline below, units 2 and 4 exhibit all of the features listed above, and other key features indicative of massive ice, such as expanded secondaries, moderately high thermal inertia in both day and night, and deep subsurface reflections with SHARAD. Thus, units 2 and 4 are interpreted to be lobate debris aprons surrounding unit 3, interpreted to be exposed massifs (Fig. 2.10).

Unit 2 contains surface textures such as polygonal terrain, expanded secondaries, moats, slope-parallel ridges, other parallel linear forms, and cracks (Figs. 2.10c, 2.11, and A2.3). Blocks, moats, and linear slope-parallel ridges can be found at the contact between units 2 and 3, interpreted to be the massif/apron margin of an LDA (Fig. 2.11). Pierce and Crown (2003) suggest moats are either secondary flow features, formed by the downhill movement of debris away from the massif, or deflation features, formed by the sublimation of ice in contact with sun-facing massifs that absorb heat. Head et al. (2010) refer to moats as shallow depressions and suggest they are also the result of sublimation of snow and ice. Moats on unit 2 do not appear to be limited to southward facing slopes. However, solar radiation likely targets all slopes in the northern summer when solar incidence is closer to 0°. Slope-parallel grooves and ridges are common in unit 2 and are suggested to be the result of massif debris flowing onto and down the aprons (e.g., Pierce and Crown, 2003), which is supported by the presence of blocks at the massif/apron margins on unit 2. Other parallel linear forms can be found on unit 2 that are not always slope-parallel (Fig. A2.3a). These are interpreted to be linear bedforms based on their larger size, inconsistent orientation, and widespread occurrence over many units in Erebus.
Unit 3 is characterized by its elevated rocky outcrops, with rounded to flat tops and blocky surfaces (Figs. 2.10, 2.11, and A2.4). Many of these massifs are rounded and symmetrical (i.e., Fig. 2.10a); however, some (~4% of unit 3) are asymmetrical with a gentle incline and steep cliff, similar to a stoss and lee morphology (i.e., Fig. 2.11). Another uncommon, but notable, feature of unit 3 is the presence of linear grooves and ridges, highlighted by the illumination angle, that trend N-S on the summits of some of the massifs (Fig. 2.11 and A2.4d). These grooved massifs only make up ~2% of unit 3 and can only be found on large massifs. Banks et al. (2008) suggests these massif lineations are grooves that are the result of wet-based glacial erosion and most analogous to terrestrial glacial grooves that indicates the direction of paleo ice flow (e.g., see Banks et al., 2008 and references therein). If this were the case, then the stoss and lee morphology of the massifs could be interpreted as streamlined hills, commonly referred to as roches moutonneés on Earth, formed from the abrasive forces on the stoss side and glacial plucking on the lee side (e.g., Glasser, 2002). The presence of lineations indicates wet-based glacial erosion, but the small percentage of occurrence indicates localized basal melting. Another possible explanation for these grooves and scarps is a tectonic origin. However, most of the massifs in unit 3 do not show this morphology making their origin indeterminate.

Unit 4 exhibits other LDA-related surface textures described by Pierce and Crown (2003), including flow-deformed ridges/rims and pits, ridge-and-valley texture, and sharp-ridge texture (Fig. 2.10d). Crenulated/brain terrain is also a common surface texture found on unit 4, which is a surface morphology that has been suggested to indicate near-subsurface massive glacial ice (Levy et al., 2009). The deformed ridges are near-concentric, sinuous, and suggest a radial direction of flow (Fig. 2.10d). Unit 2 exhibits far fewer deformed features and is much smoother (Fig. 2.10c) than unit 4; however, morphology, expanded secondaries, moderately high thermal inertia, and SHARAD subsurface reflections suggest both units are ice-rich. This difference of morphology has been documented in other LDA where the distal parts of the aprons are suggested to have experienced more sublimation and debris accumulates (Head et al., 2010). Additionally, compressional ridges are typically more prominent as ice thins and flow velocities decreases (e.g., Potter, 1972).
Deep SHARAD reflections are mostly absent for units 2–4, but what is present suggests that units 2 and 3 have subsurface reflections reaching up to 100 m depth (Table 2.2). This is likely a by-product of pixel size overlapping multiple units. Unit 3 is interpreted as massif bedrock and we do not expect subsurface ice to underly this unit. Aside from unit 3, the deepest subsurface radar reflections in Erebus are found in unit 2 (Table 2.2), which is expected based on typical SHARAD observations of LDA (e.g., Holt et al., 2008).

Units 2, 3 and 4 in Arcadia Planitia display all characteristics of LDAs which are interpreted to be debris covered glaciers that formed from the accumulation of ice and/or snow. We interpret unit 3 to be massifs of exposed bedrock, unit 2 as a potentially thicker portion of a debris covered glacier compared to its lower flanks, and unit 4 as a potentially thinner portion of a debris covered glacier, or an ice-cored moraine after the ablation of the lower flanks of the debris-covered glacier.

2.5.2.2. Possible Sources of Ice

The most widely accepted formation mechanism for LDA is through the deposition and accumulation of snow and ice in alcoves, along flanks of valley walls, or around massifs followed by the outward flow of glacial ice (Head et al., 2010). As discussed for unit 1, many studies have suggested the deposition of snow and ice in Arcadia and is the likely source for massive ice in units 2 and 4. It is unclear whether units 1, 2 and 4 formed at the same time. Units 2 and 4 have deeper subsurface reflections suggesting the ice layer is thicker in these units than unit 1. LDA have been suggested to be on the order of 100s of millions of years old (Levy et al., 2014), whereas plains ice, specifically in Arcadia, is suggested to be 10s of millions of years old (Viola et al., 2015). This is attributed to LDA ice being much thicker than plains ice based on SHARAD. Although limited SHARAD data is provided for the LDA in eastern Erebus, we assume the LDA (units 2 and 4) in the region are thicker and older than unit 1 based on other LDA age estimates (e.g., Levy et al., 2014). The LDA likely grew over time as snow accumulated during periods of high obliquity. The ice sheet that is suggested to have covered the plains of Arcadia, which
now contains unit 1, may have once blanketed the pre-existing LDA in eastern Erebus. Thus, units 2 and 4 may consist in part of remnants of ice of the same age as unit 1.

The presence of a lag deposit or the LDM over the massive ice could at least partially protect and preserve subsurface ice. The preservation of ice over these timescales are not fully understood, but the deposition of a surficial dust-ice mixture, such as the LDM, has been suggested to have had the potential to preserve massive ice units (Fastook et al., 2014).

2.5.3. Units 5 and 6: Ice-rich Mantling Units in Erebus Montes

The origin of units 5 and 6 are more difficult to discern. They most commonly occur as transitional/connecting units between other units. However, they have some distinct characteristics making these worthy of distinction.

Unit 5 is the most difficult unit to constrain as it shares many characteristics with unit 4, such as parallel linear forms, bumpy and crenulated terrain, and expanded secondaries. The parallel linear forms are likely bedforms as they commonly cross-cut units. The key feature used to differentiate the two units is the presence/absence of deformed ridges (Fig. 2.10 and A2.6c). Unit 5 is most clear in THEMIS daytime IR as it appears thermally dark and acts as an island for the Erebus region (Fig. 2.1b). This “island” is also slightly elevated compared to the surrounding terrain, appearing to mantle the region (Fig. 2.4). Unit 5 is likely an extension of unit 4 but with the absence of obvious flow features. Therefore, unit 5 is interpreted to be an ice-rich unit mantling the surrounding terrain. This is further supported by SHARAD subsurface reflections which indicates an average ice-rich layer thickness of 42 m, which is slightly thinner than the average depths reported for unit 4.

Unit 6 acts as a transitional unit between unit 1 and the Erebus “island.” Its most distinct features are the abundance of expanded secondaries and a low relative brightness. It has a similar relative brightness, smooth and polygonised morphology, and high thermal inertia as units 1 and 2. It is smoother and relatively brighter than unit 1 as it lacks the ridges
present on unit 1 (Fig. A2.7a). Deep SHARAD reflectors indicate comparable ice-rich layer thicknesses as unit 1. Unit 6 appears to overlap other units, such as units 2 and 4 (Fig. 2.10), and the surrounding terrain (Fig. A2.7d). Unit 6 is not commonly observed overlying other units as it does in Figure 2.10 making it difficult to make a concrete interpretation. Unit 6 appears to be disintegrating in patches in some areas (Fig. A2.7d) and may be explained as a retreating/ablating ice-rich blanket, possibly similar to the LDM. However, the amount of ice may vary as it can appear to act almost as a tributary to unit 1 (Fig. A2.7a). Thus, we interpret unit 6 to be an ice-rich transitional unit connecting unit 1 to the Erebus “island.”

2.6. Summary and Conclusions

We have documented the existence of viscous flow features in a flat-lying region in the lower mid-latitudes of Arcadia Planitia. We have mapped the occurrence of conventional and non-conventional VFFs, including lobate debris aprons, icy mantling units, and sinuous features resembling lineated valley fill. Three sinuous features trending NW-SE in our mapping area in Arcadia Planitia are characterized by their sinuous nature, distinct boundaries to adjacent terrain, arcuate and linear flow bands, bright thermal reflectance, moderately high thermal inertias, and deep radar reflectors and dielectric constants indicative of ice. Based on our analyses we can answer the following questions:

1) Are the sinuous features LVF? We conclude that the sinuous features are not LVF, but are instead a non-conventional VFF. We interpret the sinuous features in Arcadia Planitia to be most analogous to the pure ice streams of Siple Coast in Antarctica with a more significant topographic focusing aspect involved in the initiation and flow of ice. Other possible origins for the sinuous features, such as eskers, channelized lava flows, and frozen fluvial systems, were considered but do not account for the properties observed for these features.

2) How could snow and/or ice accumulate and flow in flat topography? The deposition and accumulation of significant snow and ice would have occurred during multiple periods of high obliquity to form an extensive ice sheet in Arcadia Planitia. The initiation of localized ice flow can occur in the form of ice streams.
The initiation and mechanism of flow is unclear but is likely due to a combination of internal ice deformation and basal slip based on terrestrial studies of ice stream flow dynamics. Further work is needed to adequately address this topic.

3) Are the sinuous features the same age as the local lobate features? The indication of widespread ice in Arcadia, in addition to the interpretation of two ice-rich units (units 5 and 6) connecting the sinuous features to the Erebus region, suggests that all the units share at least one period of ice deposition. However, LDAs tend to be thicker and older than plains ice across the mid-latitudes of Mars (e.g., Bramson et al., 2015; Levy et al., 2014; Viola et al., 2015). Thus, we suggest LDA (units 2 and 4) began to form before unit 1.

4) What is the relationship between the lobate and sinuous features? We suggest that the LDA (units 2 and 4) were deposited first and continued to grow during the deposition of unit 1 as part of the same ice sheet that collapsed and was subsequently buried. The lobate and sinuous features are connected by units 5 and 6. Further work is needed to constrain ice depositional events in Arcadia Planitia.

5) Why is eastern Erebus Montes thermally different than the sinuous features? The exposure of bedrock in the region has likely led to increased sediment deposition in the area possibly yielding a “dirty” ice and thus affecting the thermal properties of the region.

In closing, the presence of ice streams on Mars may significantly change our understanding of glacial dynamics and landsystem evolution. The use of other data products to assist in the identification of non-conventional viscous flow features is suggested for future studies on subsurface ice identification. Our mapping area in Arcadia Planitia is a flat-lying plain in the lower mid-latitudes with minimal boulders and abundant evidence for large volumes of near-surface ice, making it a potentially favorable site for future in situ resource utilization and human missions.

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Chapter 3: Vermicular Ridge Features on Dundas Harbour, Devon Island, Nunavut

3.1. Introduction

While the Canadian High Arctic has been the subject of periglacial and glacial studies in the past (e.g., England et al., 2006; French, 2017 and references therein; Mackay and Burn, 2002; Washburn, 1979) Tallurutit (Fig. 3.1a), known as Devon Island, is the largest uninhabited island in the world and is largely understudied. Devon Island is located in Nunavut and lies just north of Tallurutiup Imanga (Parry Channel). It is located along the proposed meridional limit of the extensive Inuitian Ice Sheet, near where the Laurentide Ice Sheet is proposed to have coalesced (Dyke, 1999; Dyke et al., 2002; England et al., 2006). Deglaciation of Devon Island began ~10 ka BP and exposed much of the western portion of the island by ~8 ka BP (Dyke 1999). The Devon Ice Cap (3,980 km$^3$) (Dowdeswell et al., 2004) is a glacial remnant that covers much of the eastern portion of the island (Fig. 3.1a). However, minimal evidence of depositional glacial landforms, such as eskers and moraines, and erosional striations has been observed on Devon Island (Dyke, 1999; Fortier et al., 1963; Grau Galofre et al., 2018). Additionally, Devon Island is located in the continuous permafrost zone (Brown, 1978; Zhang et al., 1999) and undergoes abundant periglacial reworking making it even more difficult to characterize the glacial history of the island (Dyke, 1999).

While investigating glacial moraines along the south coast of Devon Island near Dundas Harbour (Fig. 3.1), we observed a region characterized by unusual features with a circular to anastomosing worm-like morphology that we refer to here as Vermicular Ridge Features (VRFs). We use this term as a purely morphologically descriptive term independent of genesis, location, and sedimentology.

Ridges with circular and anastomosing morphologies are a common product of periglacial and glacial processes. Periglacial processes can produce circular features, such as sorted (Hallet and Prestrud, 1986) and non-sorted circles (Washburn, 1997), collapsed pingos (Mackay, 1998), and string bogs (Drury, 1956; Scheffers et al., 2015). One of the most common periglacial landforms are sorted circles, also referred to as stone circles,
consisting of raised ridges of course grains surrounding a central domain of fines. There are many reports of stone circles occurring in the Arctic (e.g., Washburn, 1979), including on Devon Island (Fiorenzo et al., 2006), with the most striking examples occurring in Spitsbergen, Svalbard (Hallet, 2013).

Glacial processes can also produce ridges with a circular, sinuous, or anastomosing morphology in moraines (Mollard, 2000). However, these glacial by-products vary widely in morphology, scale, material, sorting, and genesis making it difficult to classify and categorize these landforms. Many examples in western Canada consist of slightly raised ridges of silt and clay till with individual rings spanning 100s of meters (Mollard, 2000). Landforms referred to as Ice Disintegration Features in southern Alberta and Saskatchewan, Canada (Gravenor and Kupsch, 1959) resemble the morphology of Veiki moraines in northern Sweden (Hoppe, 1952; Lagerbäck, 1988), Pulju moraines in Finland (Sutinen et al., 2014), ring-ridge hummocky moraines in northern Finland (Aatolahti, 1974), and circular ridges in Norway (Knudsen et al., 2006). However, these landforms differ in morphology in terms of size, shape, relief, and grain size, and the origin of these landforms is actively debated.

Here, we report on the first analysis of VRFs on the south coast of Devon Island. We provide a comparison to previously documented circular and anastomosing ridge features produced by periglacial and glacial processes and we find these landforms to be unlike any previously reported landforms.

3.2. Geologic Setting

Dundas Harbour is located on the southeastern coast of Devon Island (Fig. 3.1). In this region, a Precambrian (~2.5 Ga) crystalline basement (Frisch, 1988) is overlain by Paleozoic sedimentary rocks (Thorsteinsson and Mayr, 1987) and unconsolidated Quaternary deposits. The western portion of Devon Island is largely dominated by Paleozoic sedimentary rocks (Harrison et al., 2015). The eastern portion is still predominantly covered by ice with Precambrian basement exposed around the island’s coastal extremities (Harrison et al., 2015).
The Devon Ice Cap (Fig. 3.1) is located a few km north of Dundas Harbour. Glacial retreat began 10 ka BP and reached its currently glaciated state ~8 ka BP (Dyke, 1998, 1999). Isostatic rebound has since led to the rising of beach sediments around much of the perimeter of the island (Dyke, 2001). Radiocarbon dates have been determined < 50 km west of Dundas Harbour at Cape Home on whale bones suggesting raised beach deposits at 11.75 m elevation were exposed ~8.73 ± 1.90 ka BP (Dyke, 2001). Raised beach deposits at the study site are at an average elevation of 11 m and were, therefore, likely exposed around the same time. The maximum Holocene marine limit near the study site reached 26 m (Dyke, 1998) and the marine limit at the study site can be located on the surficial geological materials map by Dyke (2001). The minimum age for the marine limit (37.5 m) at Cape Home is 9.92 ± 1.8 ka BP (Dyke, 1999).

Figure 3.1. Study site location observed using World Imagery (Esri, 2018). (a) Tallurutit, also known as Devon Island (75°5’33.532”N, 87°1’1.67”W), is located in Nunavut, Canada, in the Canadian Arctic Archipelago. It is the largest uninhabited island in the world with one of the largest ice caps, the Devon Ice Cap (DIC), from the last glacial maximum in the Arctic. Figure 3.1b is identified by the black box. (b) Dundas Harbour (74°31’53.072”N, 82°23’0.139”W) is located along the southeast coast of Devon Island. Valley and piedmont glaciers deposit large amount of sediment along the southeastern coast of Devon Island. The study site lies just east of Dundas Harbour and is identified by the red box. World Imagery Sources: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community.
Devon Island has limited meteorological records; however, long term climate stations are located at Qausuittuq (Resolute Bay) on Cornwallis Island (~ 375 km west of the study site) and Mittimatalik (Pond Inlet) on Qikiqtaaluk (Baffin Island) (~ 245 km south of the study site). The Resolute station data reports a mean annual air temperature of −15.7°C and mean annual precipitation of 161.2 mm (mostly in the form of snow) between 1981 and 2010 (Environment Canada, 2021a). The Pond Inlet station data reports a mean annual air temperature of −14.6°C and mean annual precipitation of 189 mm (mostly in the form of snow) between 1981 and 2010 (Environment Canada, 2021b). Climate data has been collected at Orbiter Lake, located inland on the western portion of Devon Island, in 2008 with a mean annual air temperature of −16.7°C (Godin et al., 2018). Additionally, borehole temperature measurements taken from eastern Devon Island suggest the island has a continuous permafrost thickness of 500–600 m (Smith and Burgess, 2002). Dundas Harbour is located in Subzone C of the Circumpolar Arctic Vegetation Map (CAVM Team, 2003) within the Graminoid, prostrate dwarf-shrub, and forb tundra (G2 – moist).

The surficial geological materials identified by Dyke (2001) at Dundas Harbour include beach sediments (Mr), end moraines with buried glacial ice and ice wedge polygons (Tmp), alluvial fans (Af), till veneer (Tv), ice scoured rock (Rs), non-scoured rock (Rr) and cliffs lined by talus from the exposed rocky surfaces (Rc). We created a slightly modified map of surface geologic materials was created at our study site based on field observations (Fig. 3.2).

### 3.3. Methodology

Fieldwork was conducted in July 2018 in the Dundas Harbour region. Uncrewed Aerial Vehicle (UAV) images were taken 20–30 m above the ground using a DJI Phantom 3 drone. Images are 12 megapixels with up to 1 cm/pixel resolution captured using the Pix4Dcapture application. Orthomosaics and digital elevation models (DEMs) were created using Agisoft Photoscan Professional 1.4.4. Points were extracted from each photo at full resolution to build orthomosaics and ultra-high quality dense point clouds. A maximum key point limit of 80,000 was used to extract the maximum points from a 12-megapixel image and an unlimited tie point limit was used to identify the maximum
overlapping points in all overlapping images. No depth filtering was applied to avoid the removal of any outlier points. Digital elevation models (DEMs) were then built and interpolated using the dense point clouds with a WGS 1984 UTM 16N projection.

The vertical resolution of overlapping UAV DEMs was offset by up to 145 m. To minimize this offset, DEMs were calibrated using ArcticDEM (Release 7) as a true value. ArcticDEM (Porter et al., 2018) has a 2 m resolution with absolute accuracy of about 4 m in the horizontal and vertical planes. The ArcticDEM layer consists of a collection of elevation data from DigitalGlobe satellite imagery. UAV DEM elevations were subtracted from the overlying Arctic DEM. An average difference between DEM datasets was then subtracted from the original UAV DEM to create a corrected DEM. After normalizing all DEMs to ArcticDEM, the vertical resolution of overlapping UAV DEMs is offset up to 2.5 m.

Three pits were dug into an isolated closed cell VRF: on the ridge and in the adjacent low-lying terrain on either side. Pits were dug to search for the presence or absence of near-surface massive ice, grain sorting, bedding, and/or grain imbrication, as well as to measure the thaw depth (early July 2018). A permafrost probe (a handheld metal rod) was used to measure thaw depth where additional digging was not possible. Sediment samples were collected from each pit from the diamicton layer. Sample bias exists among grains larger than 15 mm (≥ pebble) as some large grains were removed from the sample during field sampling. The largest grain observed in each pit was measured and reported. Samples were shipped to Western University for dry sieve analysis.

Sensor and Software’s PulseEKKO bistatic ground penetrating radar (GPR) system was used to search for massive ice and diamicton thickness with 200 MHz antennae. Signal velocity was calibrated based on sedimentology and thaw depth collected in the field. GPR signals penetrated to approximately 6 m; however, results yielded no clear evidence of massive ice or underlying bedrock, possibly due to abundant signal scattering from cobbles and boulders in the diamicton. Therefore, GPR data is not presented or used in this study.
Field reconnaissance was carried out between the eastern coast of Dundas Harbour and the nearest glacier to the east of the mapped study site and extended inland up onto the Precambrian bedrock plateaus. The field team compiled a bedrock (not reported here) of the region and a surficial geology map of the study site.

The sediment samples were prepared and tested for sieve analysis following the American Society for Testing Materials (ASTM) C136/C136M-19 standard for fine and coarse aggregates (ASTM Standard, 2014) to determine grading of materials. Samples were baked at 110°C for 48 hours using a Precision Thelco lab oven and the dry samples were sieved in small batches. Material was weighed to an accuracy of 0.0001 g. Sieve sizes used were based on availability and include sieve numbers 4, 5, 7, 10, 14, 16, 18, 20, 25, 30, 35, 40, 60, 100, 120 and 200. Pebbles and cobbles in the sample were removed before sieving and were weighed and measured along their three major axes. Grain size was converted to phi units (a logarithmic scale converting mm to whole integers) to analyse size frequency distributions of sediments (Krumbein, 1938).

### 3.4. Results

#### 3.4.1. Context and setting of the VRFs

The region where we identified VRFs is characterized by three terraces (Fig. 3.2), with each terrace cliff base roughly outlined by bright white snowpack in Figures 3.2a–b. Terrace 1 (T1) is the uppermost terrace with an average elevation of ~80 m and continues to rise in elevation northward into plateaus of Precambrian bedrock (Fig. 3.2c). T1 is characterized by a mixture of diamicton and blocky mass wasted debris and talus that drape frequent rocky outcrops of underlying Precambrian bedrock (Fig. 3.2c). Plateaus of Precambrian bedrock (Rr by Dyke, 2001) are prominent on T1 resulting in talus-lined cliffs (Rc by Dyke, 2001) that contribute abundant mass waste onto the terrace (Fig. 3.2c). Deep channels incise the underlying Precambrian bedrock, some of which have left alluvial fan deposits (Af by Dyke, 2001) of large blocky boulders on the surface (Fig. 3.2c). Active sorting and solifluction can be observed on T1 in regions where diamicton, interpreted as a veneer of till (Tv) by Dyke (2001), and mass wasted debris are more
prominent. Underlying Precambrian bedrock sometimes displays evidence of glacial scour (Rs by Dyke, 2001).

Terrace 2 (T2) is at a lower elevation than T1 with an average elevation of 35 m and characterized by a layer of diamicton (thickness varies across the field site) overlying bedrock (Fig. 3.3) and raised-beach deposits from T3 (Fig. 3.2d). The southern edge of T2 (~ 25 m) marks the Holocene marine limit, which is suggested to be 26 m by Dyke (1998). Beach deposits are observed to onlap the diamicton at the southernmost and westernmost edges of T2 at/near the marine limit. This terrace deposit is interpreted to be composed of glacial end moraines with buried glacial ice (Tmp) by Dyke (2001). Our field observations agree with an interpretation of glacial till moraines. No buried massive ice was exposed nor observed, but recent thermokarst activity indicates ground ice exists and has been subject to thaw. Minor solifluction, sorting and thermal contraction cracking (Figs. 3.3 and 3.4) occur in T2, consistent with Dyke’s (2001) observations. Vegetation, consisting of native black and white lichens, mosses, grasses, forbs, and shrubs, is present across the study site but is most prominent in T2 (e.g., Fig. 3.2d).

Terrace 3 (T3) is at a lower elevation than T2 with an average elevation of 11 m and is characterized by raised beach sand deposits (Mr by Dyke, 2001) that form a tombolo (Figs. 3.2a–c). T3 overlies Precambrian bedrock (Fig. 3.2c) and shallow marine deposits. Orthogonal thermal contraction polygons with an average diameter of ~40 m occur within T3.
Figure 3.2. Study site surface geologic materials. (a) The study site (74°31’35.75” N, 82°19’59.838” W) observed using World Imagery (Esri, 2018) is located just east of
Dundas Harbour and extends into Lancaster Sound along a tombolo. Three distinct terraces are present at the region and are roughly outlined by bright white snowbanks resting at the scarp base of each terrace. Five deeply incised channels are identified with red arrows. Perspectives of field photos presented in panels c and d are indicated by white arrows. (b) Map of three terraces at the study site. The uppermost terrace (T1) comprises predominantly Precambrian bedrock overlain by patchy diamicton and mass wasted debris. The middle terrace (T2) is composed predominantly of diamicton. The lowermost terrace (T3) is interpreted to be raised beach deposits. An alluvial fan (Af) of mass waste overlies all units. Hashed regions indicate regions of Vermicular Ridge Features (VRFs). Three sites of VRFs were investigated (numbered in red). (c) Field perspective of the three terraces at the study site roughly outlined. T1 includes uplifted plateaus of Precambrian bedrock in the background. The same five incised channels (red arrows) can be seen from this field perspective in T1 exposing underlying bedrock. (d) Field perspective from T3 looking up slope to T2 at Site 1 and T1. VRFs are only found on T2. (e) Field perspective of VRFs on T2 at Site 1 with exposed bedrock from T1 in the distance behind some fog. World Imagery Sources: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community.

VRFs occur on T2 exclusively along ~2‒3 km of the coast (Fig. 3.2b). The lowest elevation at which VRFs occur is ~ 25 m, around the marine limit. Three VRF sites were investigated (Figs. 3.2b, 3.3, and 3.4). Shrubs, forbs, grasses, and lichens dominate Site 1 (Figs. 3.2d‒e and 3.3) and a thin (10‒20 cm) layer of poorly developed soil is forming at the surface. Topographic lows tend to be more densely vegetated compared to the ridges (Fig. 3.2e). A distinct difference in species is evident between the ridges and the troughs of closed VRF cells at this site in particular. Dryas integrifolia, a yellow/white flowering dwarf shrub, occurs almost exclusively on the ridges, whereas Saxifraga oppositifolia, is a purple flowering forb, occurs in the central troughs of VRFs and other topographic lows. Narrow and shallow thermal contraction cracks cross-cut VRFs and form no distinct polygonal shape (Fig. 3.3). The diamicton layer appears to thicken downslope,
with an observed thickness of 2.5 m where underlying Precambrian bedrock is exposed in one location (Fig. 3.2a) upslope, and a possible maximum thickness of ~12 m downslope at the terrace cliff (Fig. 3.2d). Site 1 VRFs occur on a gentle regional slope of 4° dipping southward (Figs. 3.2e and 3.3b).

Site 2 (Figs. 3.2b and 3.4) is just west of Site 1 and is an extension of the same deposit. Grasses and mosses dominate Site 2 which is poorly drained, exhibiting standing water in the topographic lows of VRFs and thermal contraction cracks. Slumping and subsequent cracking of material due to recent thermokarst processes were observed in depressions (Fig. A3.1a). VRFs occur on a gentle slope of 3.5° dipping southwest. Ridges are less distinct and appear warped compared to those at Site 1 (Fig. 3.4). Some ridges can only be clearly identified in the DEM (Fig. 3.4b). Beach sediments appear to onlap diamicton from the west and southernmost extent of the terrace where orthogonal thermal contraction cracks begin to appear (Fig. 3.4a).

Site 3 (Figs. 3.2b and 3.4) is ~350 m east of Sites 1 and 2 and lies south of multiple deeply incised channels (Figs. 3.2a–b). The vegetation at Site 3 is dominated by lichens and mosses which, along with weathered grains at the surface, give Site 3 a reddish orange appearance (Fig. 3.4c). Standing water occurs in some of the topographic lows of VRFs and vegetation is denser within topographic lows. Beach sediments onlap the southernmost edge of the terrace. Thermal contraction cracks appear roughly orthogonal within the onlapping beach sediments and less distinct in the diamicton (Fig. 3.4c). Site 3 occurs on a gentle regional slope of 3.6° dipping southward (Fig. 3.4c). The thickest exposure of diamicton observed reaches 1.7 m at the southernmost extent of the terrace (Fig. 3.4c).

3.4.2. Morphologic Description of the VRFs

VRFs found at Dundas Harbour comprise a series of ridges and troughs that range from almost circular in planform to anastomosing ridges and troughs that are morphologically vermicular in planform (Figs. 3.2e, 3.3, and 3.4). The ridges of the VRFs are connected within the deposit by a surrounding low-lying terrain, referred to here as the mesh of the
deposit (Fig. 3.3c). Some circular to semi-circular closed cells exhibit raised convex ridges encircling a central concave trough (Figs. 3.3–3.6). Small raised circular ridges, or hummocks, without a central trough are also common forms (Fig. 3.3c). Additionally, many ridges do not make complete closed cells and have a more sinuous morphology (Fig. 3.4c). Ridges also commonly share borders (Fig. 3.3c) and appear more brain-like in morphology. It is this variability that led us to use the term vermicular, which acts as a morphologic descriptor for the circular, semi-circular, sinuous, elongated, and anastomosing (or brain-like) ridge morphologies observed in the deposit.

In closed cells, central concave troughs can be deeper (Fig. 3.5) or higher (Fig. 3.6) than the mesh, but ridges are always higher in elevation than the adjacent mesh (Fig. 3.3c). Occasionally, small piles of cobbles (Fig. 3.5) and ponds of water are present in the topographic lows of the troughs. One closed cell has exposed underlying bedrock in its central trough (Fig. 3.3). The transition from ridge to central trough of a closed cell is gradual and bowl-shaped (Figs. 3.5 and 3.6), whereas the transition from ridge to its adjacent mesh has steeper slopes in most cases (Fig. 3.5).
Figure 3.3. Vermicular Ridge Features (VRFs) at Site 1. (a) UAV orthomosaic of Site 1. Green arrows point to thermal contraction cracks. Underlying Precambrian bedrock can be found outcropping at the yellow arrow. Figure 3.5 location identified with white arrow. Longest closed cell (Fig. A3.2a) identified by red arrow. (b) UAV DEM of VRFs at Site 1. Elevation in meters. Ridges form circular, sinuous, and anastomosing forms surrounded by a low-lying mesh. Closed cells contain low-lying central depressions. (c) 3D perspective of Site 1 with 2x vertical exaggeration looking uphill to the north-northwest. a and b – Crack along the axial trace of a ridge. c – Low-lying terrain within the diamicton deposit, referred to as mesh, surrounding and connecting VRFs. d – Small
circular VRF without a central trough. e – One example of ridges sharing borders creating a series of interconnected ridges. f – Thermal contraction crack cross-cutting ridges. g – Underlying Precambrian bedrock outcropping in the center of a closed semi-circular cell.

Figure 3.4. Vermicular Ridge Features (VRFs) at Sites 2 and 3. (a) UAV orthomosaic of Site 2 located just west of Site 1. VRFs appear warped and become increasingly misshapen and poorly defined downslope. Dark areas on the west side of the image are ponds of water. Sandy beach sediment on the west side of the image onlaps and underlies the diamicton deposit at this site. Green arrows point to thermal contraction cracks. (b) DEM of Site 2 displays a lower relief between ridges and low-lying mesh and central depressions compared to Site 1. Elevation in meters. (c) UAV orthomosaic of Site 3 located east of Sites 1 and 2. Lichens, mosses and weathered grains give the surface a
red/orange appearance. Dark areas indicate ponds of water and/or increased vegetation. Bright white areas are patches of snow. Green arrows point to thermal contraction cracks. Red arrows indicate where the VRF deposit is topographically higher than the surrounding onlapping sandy beach deposits and indicate the location of figure 3.6. Longest ridge example at this site can be seen in Figure A3.2b. (d) DEM of Site 3 showing a series of circular to anastomosing ridges. Elongated ridge runs perpendicular to slope. Elevation in meters.

Ridges reach up to 2.5 m in height when measuring from the lowest point in the central trough to the apex of the ridge. However, only one example of a 2.5 m relief ridge is observed in a closed cell where underlying bedrock outcrops (Fig. 3.3). Ridge height generally ranges between 0.5 to 1 m. Ridge width ranges between 1.5 and 6 m. The diameter of closed cells is variable with the longest examples oriented perpendicular to slope (Fig. 3.3a). Closed cells with a central trough have diameters ranging from 4 to 12 m from ridge-to-ridge (Figs. 3.5 and 3.6). Circular ridges without a central trough are less than 4 m in diameter (Fig. 3.3c). Spacing between ridges ranges from 1 to 10 m with some instances of isolated circular ridges as far as 30 m away.

The longest spanning closed cell occurs at Site 1 and spans 72 m (Figs. 3.3a and A3.2a). The longest spanning ridge occurs at Site 3 and spans a 120 m distance (Figs. 3.4c and A3.2b). This ridge marks the end of the deposit where it appears to overlie raised beach deposits. Elongated ridges commonly extend perpendicular to slope; however, do not appear to occur on slopes steeper than the average of the deposit (> 3.5–4°).

Many ridges exhibit a crack running along or just off-centre from the axial trace of the ridge crests (Figs. 3.3c and A3.1b). Where ridges are vegetated, splitting of the soil and vegetation at the axial crack can be observed (Fig. A3.1). Shallow and narrow thermal contraction cracks exist in T2 and crosscut the ridges (Figs. 3.3 and 3.4) but do not form a polygonal network.
Figure 3.5. Topographic profile from A to A’ (14 m across) of a closed semi-circular cell in Site 1 (Fig. 3.3a). Convex ridge reaches 1.13 m in height. Ridge to Ridge spans 10 m. The transition from deposit mesh to ridge is a sharp elevation change compared to gradual elevation change from ridge to central trough. Transition from ridge to center of trough is bowl-shaped. Piles of cobbles rest within the trough and can be seen in the topographic profile. Three pits (Fig. 3.7) were dug to the frost table on July 4, 2018. The thaw depth on that date is interpolated by the dotted line.

Figure 3.6. Topographic profile from B to B’ (17.5 m across) of a closed circular cell in Site 3 (Fig. 3.4c). Convex ridges have slight disruptions where cracks occur near the apex of the ridges. Ridges gradually transition into a concave central trough that lies higher in elevation than the terrain outside of the closed cell.

3.4.3. Sedimentology

Three pits about 50 cm wide were excavated to depths of 48‒70 cm at a closed semi-circular cell at Site 1 (Figs. 3.3, 3.5, and 3.7). This cell has an asymmetrical topographic
profile with a maximum ridge height of 113 cm (Fig. 3.5). Sorting, bedding, grading, grain imbrication and massive ice were not observed in the pits.

Pit 1 was dug in the center of the circular trough from a vegetated surface to the thaw depth at 48 cm (Figs. 3.5 and 3.7). A 10 cm layer of darker-colored organic-rich sandy soil with roots overlies 38 cm of a relatively lighter-colored diamicton (Fig. 3.7). The largest grain size identified in the pit is a cobble with a 21.5 cm long axis. Pebbles and cobbles reside in a predominantly sandy matrix (Fig. 3.8). The substrate was moist in comparison to soil in Pits 2 and 3.

Pit 2 was dug at the apex of the highest ridge in the cell (Figs. 3.5 and 3.7) from a lichen-rich surface to a depth of 68 cm. A permafrost probe was used to measure an additional 51 cm to the frost table giving a thaw depth of 119 cm. Roots penetrated to a depth of 20 cm and overlie 99 cm of diamicton (Fig. 3.7). The largest grain identified in the pit was a boulder with a 42 cm long axis. Pebbles, cobbles and small boulders reside in a predominantly sandy matrix (Fig. 3.8). The substrate was drier than soil in Pits 1 and 3.

Pit 3 was dug in the adjacent terrain outside of the closed cell (Figs. 3.5 and 3.7) from a vegetated surface to a depth of 70 cm. A permafrost probe was used to measure an additional 16 cm to the thaw depth of 86 cm. Roots penetrated 13 cm and overlie 73 cm of diamicton (Fig. 3.7). The largest grain identified in the pit was a boulder with a 27 cm long axis. Pebbles, cobbles and small boulders reside in a predominantly sandy matrix (Fig. 3.8).

Material from all three pits is polymodal and dominated by pebbles and cobbles in a sandy matrix (Fig. 3.8). Pebble and cobble concentrations are higher than reported here for each pit due to sample bias from removing large grains (≥ pebble) during sampling.
Figure 3.7. Three pits were dug in a semi-circular closed cell at Site 1 (Fig. 3.2a and 3.5) on July 4, 2018. Pit 1 was dug in the central trough of the closed cell. The upper 10 cm (dotted red line) consisted of dark brown organic-rich soil overlying diamicton to a thaw depth of 48 cm. Pit 2 consists of roots penetrating through the upper 20 cm of diamicton. The pit was dug to a depth of 68 cm and a permafrost probe penetrated an additional 51 cm to the frost table. Pit 3 consists of long roots penetrating through the upper 13 cm of diamicton. The pit was dug to a depth of 70 cm and a permafrost probe penetrated an additional 16 cm to the frost table.
3.5. Discussion

We have documented a landform that we refer to as Vermicular Ridge Features (VRFs) along the coast of Dundas Harbour on southern Devon Island, Nunavut, Canada. Numerous raised-ridge features with a circular, sinuous, or anastomosing morphology have been described in the literature and have been interpreted to have formed as a result of periglacial (e.g., Fig. 3.9, Table 3.1) and/or glacial (e.g., Fig. 3.10, Table 3.2) processes. Periglacial processes can produce circular to anastomosing raised-ridge features such as sorted circles (Figs. 3.9a, b), non-sorted circles (e.g., Washburn, 1997), involuted hills (Fig. 3.9c; Mackay, 1963), collapsed pingos and palsas (Fig. 3.9d; Mackay, 1998), retrogressive creeping slumps (Fig. 3.9e; McWade et al., 2017), and string bogs (Fig. 3.9f; Drury, 1956; Scheffers et al., 2015). Glacial processes can produce a variety of circular to anastomosing ridge features (e.g., Mollard, 2000; Fig. 3.10), that may be broadly categorized as hummocky moraines (Benn and Evans, 2010; Fredin et al., 2013), but that have been referred to by a wide variety of names in the literature and that lack consensus on a formation mechanism. These circular to anastomosing periglacial and glacial ridge features range widely in morphology, scale, sedimentology, and genesis. In the following sections, we first compare the VRFs at Dundas Harbour to documented and morphologically similar periglacial and glacial ridge features in the literature, before proposing an origin.
Figure 3.9. Circular, sinuous and anastomosing raised-ridge features produced by periglacial processes observed in World Imagery (Esri, 2018). World Imagery Sources: Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community. A) Sorted stone circles in Broggerhalvoya, NW Spitsbergen (Hallet, 2013) (11°28′48.087″E, 78°56′59.752″N). B) Sorted stone circles in
Figure 3.10. Circular, sinuous and anastomosing raised-ridge features produced by glacial processes observed in World Imagery (Esri, 2018). World Imagery Sources: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGrid, IGN, and the GIS User Community. (a) Ice Disintegration Features in Starland County, Alberta, Canada (Gravenor and Kupsch, 1959) (112°13'46.363"W, 52°2'43.667"N). (b) Brain-like deformed rings West of Zenon Park, Saskatchewan, Canada in farmland.

3.5.1. Periglacial Evaluation

Stone circles (Figs. 3.9a, b, and 3.11), also referred to as sorted circles, are a common periglacial landform across the High and Low Arctic, in regions such as Spitsbergen, Canada, Alaska, and Greenland (Figs. 3.9a, b, Table 3.1) (Hallet, 2013; Hallet and Prestrud, 1986; Schmertmann and Taylor, 1965; Washburn, 1997, 1973, 1956), and are characterized by their stark contrast between their circular to semi-circular raised ridge of coarse grains compared to their central, often convex upward, domain of fine grains. Labyrinthine patterned ground, described by Kessler and Werner (2003), is a poorly documented landform that is morphologically similar to VRFs at Dundas Harbour. This landform is an irregular “labyrinthine” form of stone circles that has been modeled to form either due to a decrease in coarse grain concentration (Kessler and Werner, 2003) or an increase in slope (Hallet, 2013).
Figure 3.11. Sorted “stone” circles at Dundas Harbour, Devon Island. Stone circles of varying size and shape located on plateaus of Precambrian bedrock just northeast of the study site at Dundas Harbour (see Fig. 3.2c for location). iPhone 7 for scale (138 mm x 67 mm).

The most prominent characteristic of stone circles is the stark grain size difference between the coarse-grained ridge and fine-grained center, as reflected in the alternative name, “sorted circles” (Hallet, 2013). The transition from the gravel ridge to the central domain is defined not only by a textural difference, but also by a microtopographic depression, or moat (Hallet, 2013). Stone circles range from 2–5 m in diameter with a ridge width of 0.5–1 m and a ridge height ranging from a few millimeters to 0.5 m and form over hundreds of years (Hallet, 2013; Hallet and Prestrud, 1986; Kessler et al., 2001). Stone circles generally occur on nearly horizontal surfaces but can occur on slopes as high as 30° (Washburn, 1973). Where stone circles do occur on steep slopes, ridges are generally elongated in the downslope direction (parallel to slope) and eventually become sorted stripes (Hallet et al., 2013).

Stone circle scale is comparable to some closed-cell VRFs at Dundas Harbour; however, stone circles are ultimately smaller. At Dundas Harbour, closed-cell ridges have a closed cell diameter ranging between 4‒72 m and ridges reach up to 2.5 m in height. Additionally, VRFs lack evidence of grain sorting (Fig. 3.8) and can be seen elongated perpendicular to slope (Figs. 3.3 and 3.4) rather than parallel to slope like stone circles.

In terms of microtopography, VRFs are also not comparable to mature stone circles; however, incipient stone circles are suggested to have well-established microrelief prior to significant lateral sorting (Hallet and Prestrud, 1986). Hallet and Prestrud (1986) suggest that stone circles in Spitsbergen began forming within the last 9 ka BP. Although differential upheaving may produce microrelief, these authors suggest that the expansion, or lateral movement, of stone circles occurs at a potentially consistent rate of 10 mm/yr. This suggests the onset of stone circle formation at their study site began up to 500 years ago to produce an average stone circle diameter of 3 m. The eastern coasts of Devon
Island deglaciated approximately 10 ka BP (Dyke, 1998, 1999), which is used as an upper limit for the initiation of periglacial processes as deglaciation of Dundas Harbour may have occurred more recently. Thus, we would expect to see well-developed (i.e., well-sorted) stone circles in the region at present given the timescale based on stone circle formation models. Additionally, we see well-developed stone circles and sorted patterned ground on the plateau north of the study site (Fig. 3.11) which has deglaciated more recently than the lowlands where the Dundas VRFs are observed. This suggests enough time has passed for the formation of well-developed stone circles in the region.

It is notable that VRFs commonly exhibit cracks along the axial trace of ridges. Similar cracks can be found on stone circles which are attributed to zones of maximum soil upwelling (e.g., Kaab et al., 2014). For example, cracks can be found on the central fine domains of stone circles as well as along the gravel ridges that are likely due to frost heave and soil upwelling (Kaab et al., 2014). Polygonal cracks, likely the result of thermal contraction, can be observed at Dundas Harbour. The cracking observed along the axial trace of VRFs appears to be different from polygonal cracking as polygonal cracking cross-cuts VRFs in many locations (Figs. 3.3 and 3.4). Regardless, cracks along some VRFs appear to be indicative of recent thermokarst and surface collapse rather than a product of thermal contraction or differential upheaving (Fig. A3.1).

In summary, Dundas Harbour VRFs are comparable to stone circles mainly in environmental conditions, but are not comparable in scale, microtopography, relationship with slope, and sorting. We interpret axial cracking and polygonal cracking to be periglacial modification of a pre-existing feature rather than the result of soil upwelling which is inherently involved in the production of stone circles.

Next, we consider non-sorted circles (Table 3.1). Non-sorted circles are the most common form of patterned ground, include many sub-types, and are prevalent in the Canadian High Arctic (e.g., Washburn, 1997). They are characterized as small circular mounds that reach up to 0.5 m in height and range 0.5–3 m in diameter. They are usually composed of fine-grained soil that becomes displaced upward due to the freeze and thaw of ice lenses. However, stony earth circles are a form of non-sorted circles that consist of
coarse grains (e.g., Williams, 1959; Rissing and Thorn, 1985). Although these features are not sorted, the small scale, and microtopography (i.e., lack of central depressions and raised ridges) are inconsistent with observations of Dundas Harbour VRFs.

Other periglacial features exhibit a series of circular to anastomosing raised ridges (Table 3.1). One well-documented example is the Involuted Hill sites near and around Tuktoyaktuk, Northwest Territories, Canada which have a distinct ridged surface (Fig. 3.9c, Table 3.1) (Mackay, 1963; Rampton, 1988). Involuted Hill is a flat-topped 25 m tall hill stretching ~700 m by 1500 m. The hill consists of a 20 m thick massive ice body overlain by approximately 5 m of clay till and peat and underlain by stratified sand and gravel (Annan and Davis, 1976; Mackay and Dallimore, 1992). The overlying material exhibits a ridge and swale morphology that has been modified by frost action and ice wedge formation. Ridges appear to be ~10 to 40 m wide with a continuous ridge encircling the hill (Fig. 3.9c). The massive ice is thought to be the result of downward freezing of glacial meltwater and porewater expulsion (Mackay and Dallimore, 1992). The differential degradation of the ice is thought to produce the series of ridges and troughs at the surface. However, the involuted hills are not morphometrically consistent with VRFs.

Retrogressive creeping slumps are a newly classified landform located in the Lac de Gras region of the Northwest Territories, Canada (Fig. 3.9e, Table 3.1) (McWade et al., 2017). These features consist of curvilinear ridges with an adjacent trough and hillslope and exist in hummocky till-like diamicton deposits. Ridges have an average height of 5.8 m and reaching up to 25 m, length of 33 to 430 m and trough width of 2 to 17 m. The trough occurs at the top of or adjacent to the ridges and appears morphologically reminiscent of the axial cracks observed along VRFs at Dundas Harbour. These features overlie massive ice and are suggested to form due to the melt of ground ice and subsequent creeping and slumping of the overlying material. A study nearby at Contwoyto Lake just north of Lac de Gras suggests buried ice in this region may be buried glacial ice or sourced from glacial meltwater (Wolfe, 1998). However, retrogressive creeping slumps are not morphometrically comparable to VRFs.
String bogs (Fig. 3.9f, Table 3.1) (Drury, 1956; Scheffers et al., 2015) and other peatland permafrost features (e.g., Way et al., 2018) exhibit a series of circular to anastomosing ridges and troughs. These features consist of ridges of peat and vegetation which does not match observations of VRFs in Dundas Harbour.
Table 3.1. Circular, sinuous and anastomosing ridge landform morphometrics and characteristics compared to Dundas Harbour Vermicular Ridge Features.

<table>
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<th>Name</th>
<th>Location</th>
<th>Width</th>
<th>Diameter/Length</th>
<th>Height</th>
<th>Material</th>
<th>Formation mechanism*</th>
<th>Reference**</th>
<th>Figure</th>
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<td>1.5–6 m</td>
<td>4–120 m</td>
<td>up to 2.5 m</td>
<td>clast-rich sandy till</td>
<td>differential ablation</td>
<td>this study</td>
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<td>Sorted &quot;stone&quot; circles</td>
<td>Spitsbergen, Svalbard, Norway</td>
<td>0.5–1 m</td>
<td>2–5 m</td>
<td>few mm to 0.5 m</td>
<td>sand, silt, gravel</td>
<td>frost heave and sorting</td>
<td>Hallet, 2013</td>
<td>3.9a, b</td>
</tr>
<tr>
<td>Non-sorted &quot;plug&quot; circles</td>
<td>Canadian High Arctic</td>
<td>N/A</td>
<td>0.5–3 m</td>
<td>0.5 m</td>
<td>mud; fine grained soil</td>
<td>cryoturbation or diapiric displacement</td>
<td>Washburn, 1997</td>
<td></td>
</tr>
<tr>
<td>Involuted Hills</td>
<td>Tuktoyaktuk, Northwest Territories, Canada</td>
<td>10–30 m***</td>
<td>up to 1500 m</td>
<td>up to 5 m</td>
<td>clay-rich till</td>
<td>differential ablation; ablation; thermokarstic degredation</td>
<td>Mackay, 1963; Rampton, 1988</td>
<td>3.9c</td>
</tr>
<tr>
<td>Collapsed Pingos/Palsas</td>
<td>Eaglesham, Alberta, Canada</td>
<td>—</td>
<td>150 m</td>
<td>1 m</td>
<td>fenland</td>
<td></td>
<td>Mollard, 2000</td>
<td>3.9d</td>
</tr>
<tr>
<td>Pingo-like mounds</td>
<td>Banks Island, Northwest Territories, Canada</td>
<td>N/A</td>
<td>40–245 m</td>
<td>2.5–14 m</td>
<td>fluvial sediments</td>
<td>frozen taliks or ice segregation</td>
<td>Pissart and French, 1976, 1977</td>
<td></td>
</tr>
<tr>
<td>Pingo-like ridges</td>
<td>Banks Island, Northwest Territories, Canada</td>
<td>20–100 m</td>
<td>2 km</td>
<td>10 m</td>
<td>fluvial sediments</td>
<td>frozen taliks</td>
<td>Pissart and French, 1977</td>
<td></td>
</tr>
<tr>
<td>Retrogressive creeping slumps</td>
<td>Lac de Gras, Northwest Territories, Canada</td>
<td>—</td>
<td>33–430 m</td>
<td>5.8–25 m</td>
<td>diamicton</td>
<td>both</td>
<td>McWade et al., 2017</td>
<td>3.9e</td>
</tr>
<tr>
<td>String Bogs</td>
<td>western Russia</td>
<td>1–3 m</td>
<td>1 km</td>
<td>1 m</td>
<td>peat/fenland</td>
<td>thermokarstic degredation</td>
<td>Harris et al., 1988; Scheffers and Kelletat, 2015</td>
<td>3.9f</td>
</tr>
</tbody>
</table>

* Ridge formation mechanism
** Only reporting on these specific examples. Many of these landforms are quite common globally and experience minor variation in morphometrics from what is reported in this table.
*** Measured by authors of this study using World Imagery (Esri, 2018)
3.5.2. Glacial Evaluation

Having considered and ruled out a periglacial origin for VRFs, we now explore glacially-derived circular to anastomosing raised-ridge features in the literature to identify a potentially more morphometrically analogous landform to the VRFs at Dundas Harbour. Glacial circular to anastomosing raised-ridge features (e.g., Fig 3.10, Table 3.2) are typically found in, or described as, hummocky moraines (also referred to as “doughnuts,” “doughnut hummocks,” “rim ridges,” “rimmed kettles,” “humpies,” “prairie mounds,” and many more) and have been identified in Western Canada (Figs. 3.10a–d; Evans et al., 2014; Gravenor and Kupsch, 1959; Johnson and Clayton, 2005; Mollard, 2000; Parizek, 1969; Paulen and McClanaghan, 2014), Northern United States (Parizek, 1969 and references therein), Norway, including circular moraine features (Ebert and Kleman, 2004) and circular ridges (Fig. 3.10e; Knudsen et al., 2006), Finland, including ring ridge hummocky moraines (Fig. 3.10f; Aartolahti, 1974), also referred to as Pulju moraines (Middleton et al., 2020; Sutinen et al., 2019, 2018, 2014), Wales, such as ramparted depressions (Ross et al., 2019), and Sweden, including Veiki moraines (Hoppe, 1952; Lagerbäck, 1988) (Table 3.2). Pulju moraines, Rogen moraines, and Veiki moraines are not categorized as hummocky moraines (Benn and Evans, 2010; Menzies and Evans, 2000), but there is a type of Rogen moraine termed ‘hummocky ribbed moraine’ (Dunlop and Clark, 2006). The term ‘hummocky moraine’ has been used for a wide range of glacial landforms and is used more as a morphological indicator rather than a sedimentological term (Menzies and Evans, 2000). Hummocky moraine is used here as a purely descriptive term that Benn and Evans (2010) broadly characterized as having a “moundy, irregular morainic topography ranging 2–70 m in relief and forming chaotic assemblages of mounds ranging 15–400 m in diameter.” Therefore, we use the term in a broad morphological sense regardless of moraine origin.

The examples of circular to anastomosing raised-ridge features in glacial moraines, which we will refer to under an umbrella term of ring-ridge moraines, all share a similar morphology loosely characterised by irregular clusters of elongated to circular ridges and troughs that occur in glacial till in previously glaciated areas (Table 3.2). However, there
is a considerable variation in scale, landform association, and sedimentology, among these features (Fig. 3.10; Table 3.2). For example, many of the ring-ridge moraines found in Alberta and Saskatchewan, Canada, occur in farmland regions (Figs. 3.10a–d) with a ridge sedimentology dominated by silty lacustrine clay sediment and till (e.g., Gravenor and Kupsch, 1959; Molland, 2000 and references therein) and stretch 100s of meters in diameter with a ridge to trough relief varying from <1 m up to 12 m (Molland, 2000) (Table 3.2). In contrast, Pulju moraines in Finland (Aartolahti, 1974; Sutinen et al., 2014) and circular ridges in Norway (Knudsen et al., 2006) are composed of sandy and silty clast-rich till (Table 3.2). Pulju moraines are 20–200 m in diameter and range from 0.5–4 m in height (Aartolahti, 1974), and circular ridges in Norway are 50–100 m in diameter and range from 2.5 to 10 m in height (Knudsen et al., 2006) (Table 3.2).
Table 3.2. Ring-ridge moraine morphometrics and characteristics compared to Dundas Harbour Vermicular Ridge Features.

<table>
<thead>
<tr>
<th>Name</th>
<th>Location</th>
<th>Time of deglaciation (BP)</th>
<th>Diameter</th>
<th>Height</th>
<th>Material</th>
<th>Landform Association</th>
<th>Formation mechanism</th>
<th>Reference</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vermicular Ridge Features (VRFs)</td>
<td>Devon Island, Nunavut, Canada</td>
<td>&lt; 8 ka</td>
<td>4‒72 m</td>
<td>up to 2.5 m</td>
<td>clast-rich sandy till</td>
<td>thermal contraction</td>
<td>supraglacial</td>
<td>this study</td>
<td>3.2</td>
</tr>
<tr>
<td></td>
<td>southern Alberta, Canada</td>
<td>~ 12.3 ka; 14 ka</td>
<td>up to 15 m</td>
<td>up to 5 m</td>
<td>till, possibly clayey</td>
<td></td>
<td>supraglacial</td>
<td>Evans et al., 2014</td>
<td></td>
</tr>
<tr>
<td></td>
<td>north-central Alberta, Canada</td>
<td>~12.9 ka ±0.8 ka; 11 ka; 14 ka</td>
<td>—</td>
<td>2–10 m</td>
<td>sandy silt/silty clay with 5-10% clast till</td>
<td></td>
<td>subglacial</td>
<td>Paulen and McClenaghan, 2014</td>
<td></td>
</tr>
<tr>
<td>Hummocky terrain</td>
<td>Finland</td>
<td>9010 and 9750 ka</td>
<td>30–100 m</td>
<td>&lt; 1.5 m; 1.6–2.1 m</td>
<td>gravelly to sandy and silty till</td>
<td>faults, flutes, eskers</td>
<td>subglacial</td>
<td>Sutinen et al., 2014</td>
<td></td>
</tr>
<tr>
<td>Pulju Moraines</td>
<td>Finland</td>
<td>12,630 ka*</td>
<td>75–150 m; up to 300 m</td>
<td>6 m; up to 10 m</td>
<td>silty clay till</td>
<td></td>
<td>either</td>
<td>Gravenor and Kupsch, 1959</td>
<td>3.10a</td>
</tr>
<tr>
<td>Circular Ridges</td>
<td>Norway</td>
<td>~14 ka</td>
<td>50–100 m</td>
<td>2.5–10 m</td>
<td>clast-rich sandy till</td>
<td>drumlins, marginal moraines</td>
<td>supraglacial</td>
<td>Knudsen et al., 2006</td>
<td>3.10e</td>
</tr>
<tr>
<td>Ice Disintegration Features</td>
<td>Saskatchewan, Canada</td>
<td>40–50 ka</td>
<td>100s of m</td>
<td>6–10 m</td>
<td>clayey silt and occasional clasts - glacial diamict and glaciolacustrine seds</td>
<td></td>
<td>either</td>
<td>Ross et al., 2019</td>
<td></td>
</tr>
<tr>
<td>Ramparted Depressions</td>
<td>Wales</td>
<td>—</td>
<td>comparable to Gravenor</td>
<td>up to 7 m</td>
<td>—</td>
<td>—</td>
<td>either</td>
<td>Lagerbaeck, 1988</td>
<td></td>
</tr>
<tr>
<td>Veiki Moraines</td>
<td>Sweden</td>
<td>40–50 ka</td>
<td>100s of m</td>
<td>6–10 m</td>
<td>clay, silt, sand, gravel</td>
<td>drumlins</td>
<td>both</td>
<td>Aartolahti, 1974</td>
<td></td>
</tr>
<tr>
<td>Ring Ridge Hummocky moraines</td>
<td>northern Finland</td>
<td>—</td>
<td>20–200 m</td>
<td>0.5–4 m</td>
<td>sandy silt till with some gravel and with boulders on ridges overlying clay-rich till</td>
<td>patterned ground</td>
<td>supraglacial</td>
<td>Aartolahti, 1974</td>
<td>3.10f</td>
</tr>
</tbody>
</table>
Table 3.2 continued…

<table>
<thead>
<tr>
<th>Name</th>
<th>Location</th>
<th>Time of deglaciation (BP)</th>
<th>Diameter</th>
<th>Height</th>
<th>Material</th>
<th>Landform Association</th>
<th>Formation mechanism</th>
<th>Reference</th>
<th>Figure</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deloraine rings</td>
<td>Quebec</td>
<td>**</td>
<td>few m to 500 m; avg 100–200 m</td>
<td>1–6 m</td>
<td>silty clay till; glaciolacustrine mud</td>
<td>—</td>
<td>ice-marginal lake</td>
<td>Dionne, 1978</td>
<td></td>
</tr>
<tr>
<td>Rimmed mounds</td>
<td>Saskatchewan, Canada</td>
<td>**</td>
<td>—</td>
<td>—</td>
<td>silty clay till; lacustrine</td>
<td>once had ice marginal lakes</td>
<td>supraglacial</td>
<td>Mollard, 2000</td>
<td></td>
</tr>
<tr>
<td>Raised rim mounds</td>
<td>Saskatchewan, Canada</td>
<td>**</td>
<td>—</td>
<td>1–2 m</td>
<td>silty clay till; lacustrine; thin gravely lag occasional lacustrine silt and clay; glaciolacustrine lacustrine silt and clay till; pebbly and cobbly surface</td>
<td>once had ice marginal lakes</td>
<td>ice-marginal lake</td>
<td>Mollard, 2000</td>
<td></td>
</tr>
<tr>
<td>Brain-like pattern</td>
<td>Saskatchewan, Canada</td>
<td>**</td>
<td>—</td>
<td>fraction of a meter</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>ice-marginal lake</td>
<td>Mollard, 2000</td>
</tr>
<tr>
<td>Circular doughnuts</td>
<td>Manitoba, Canada</td>
<td>**</td>
<td>—</td>
<td>&lt; 1 m</td>
<td>—</td>
<td>once had ice marginal lakes</td>
<td>ice-marginal lake</td>
<td>Mollard, 2000</td>
<td></td>
</tr>
<tr>
<td>Doughnut rings</td>
<td>Saskatchewan, Canada</td>
<td>**</td>
<td>—</td>
<td>&quot;very low relief&quot;</td>
<td>clay-rich till</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>Mollard, 2000</td>
</tr>
<tr>
<td>Doughnut mounds</td>
<td>Saskatchewan, Canada</td>
<td>**</td>
<td>—</td>
<td>up to 12 m</td>
<td>clay-rich till</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>Mollard, 2000</td>
</tr>
<tr>
<td>Ice-Contact Rings</td>
<td>Saskatchewan, Canada</td>
<td>—</td>
<td>10s of m</td>
<td>1.5–10.5 m</td>
<td>till, sand and gravel laminated tills with interbedded sorted sediments (Kalix till and Sveg till)</td>
<td>—</td>
<td>supraglacial</td>
<td>Parizek, 1969</td>
<td></td>
</tr>
<tr>
<td>Rogen Moraine</td>
<td>Sweden</td>
<td>—</td>
<td>up to 100 m</td>
<td>up to 30 m</td>
<td>—</td>
<td>—</td>
<td>subglacial</td>
<td>Lundqvist, 1989</td>
<td></td>
</tr>
<tr>
<td>Circular moraine features (CMF)</td>
<td>northern Norway</td>
<td>11–15 ka</td>
<td>20–170 m</td>
<td>0.5–10 m</td>
<td>diamicton</td>
<td>—</td>
<td>englacial</td>
<td>Ebert and Kleman, 2004</td>
<td></td>
</tr>
</tbody>
</table>

* date from Klassen, 1993

** Not provided. Assuming 12-14 ka based on regional context
Two main formation hypotheses for ring-ridge moraines have been proposed, but their origin remains debated. The first is that they represent supraglacial and englacial debris concentrations left from the passive ablation of stagnant “dead” ice (Aartolahti, 1974; Boulton, 1972, 1967; Clayton, 1967; Clayton et al., 1985; Clayton and Moran, 1974; Ebert and Kleman, 2004; Eyles, 1983, 1979; Gravenor and Kupsch, 1959; Ham and Attig, 1996; Jennings, 2006; Johnson et al., 1995; Johnson and Clayton, 2005; Knudsen et al., 2006; Krüger, 1983; Molland, 2000; Parizek, 1969; Patterson, 1998, 1997; Paul, 1983; Paulen and McClenaghan, 2014; Ross et al., 2019; Schomacker, 2008; Sollid and Sørbel, 1988). Sandy and clast-rich ring-ridge moraines are suggested to form from the settling of supraglacial and englacial till due to spatial changes in debris thickness, and consequently, solar insulation. Clay-rich ring-ridge moraines are suggested to form due to ice blocks settling and melting in a drained proglacial or ice-marginal lake leaving predominantly fine-grained till and lacustrine sediments. Alternatively, it has been proposed that these landforms formed via subglacial diapirism and squeezing of subglacial water-saturated till into basal crevasses and cavities of a stagnant or disintegrating glacier leaving till ridge rings (Aartolahti, 1974; Boone and Eyles, 2001; Eyles et al., 1999; Gravenor and Kupsch, 1959; Hoppe, 1952; Johnson and Clayton, 2005; Menzies and Shilts, 2002; Molland, 2000; Stalker, 1960). If sand or clast-rich supraglacial till is present, a layered ridge may exist with coarse-grained till overlying fine-grained till (Figs. 3.9d–e), as has been observed in Finnish ring ridge hummocky moraines (Aartolahti, 1974).

Additional suggested formation hypotheses include subglacial meltwater erosion similar to that of drumlins and Rogen moraines, and other transverse moraines (Munro-Stasiuk and Sjogren, 2006; Munro-Stasiuk and Shaw, 1997; Shaw, 1996); proglacial blowout of over-pressurized groundwater (Bluemle, 1993; Boulton and Caban, 1995; Evans, 2009, 2003; Evans et al., 1999); and seismically induced subglacial diapirism (Middleton et al., 2020; Sutinen et al., 2019, 2018, 2014).

Overall, the Dundas Harbour VRFs are morphologically and morphometrically comparable to ring-ridge moraines (Fig. 3.10; Table 3.2). However, VRFs in Dundas
Harbour are much smaller in scale than any presently documented ring-ridge moraine. Yet, the absence of vertical or lateral sorting at our study site, in what we are interpreting to be glacial till based on sorting, roundness, grain size distribution, and depositional environment, suggests Dundas VRFs were deposited in situ as a type of hummocky moraine. A sandy clast-rich till, as observed in Dundas ridges, has only been documented for Pulju moraines (Aartolahti, 1974; Sutinen et al., 2014), circular ridges described by Knudsen et al. (2006), and rim ridges (now known as Veiki moraines) described by Hoppe (1952) and Lagerbäck (1988). This sedimentology and morphology may be a rare occurrence or not well documented.

Regardless of sedimentology and scale, it is also important to note that many of the ring-ridge moraines discussed above and listed in Table 3.2 occur in association with other glacial landforms, suggesting that despite their differences at least some of the ring-ridge moraines likely form in a similar manner to their associated landforms. For example, many ring-ridge moraines have been documented with landforms, such as flutings (Evans et al., 2014; Gravenor and Kupsch, 1959; Paulen and McClenaghan, 2014; Sutinen et al., 2014), drumlins (Gravenor and Kupsch, 1959; Hoppe, 1952; Knudsen et al., 2006; Paulen and McClenaghan, 2014) or eskers in their study sites (Evans et al., 2014; Sutinen et al., 2014) (Table 3.2). Many ring-ridge moraines are also documented to be in association with ice streams, suggesting wet-based glaciation and/or significant sliding may be involved in the formation of these features (Evans et al., 2014; Knudsen et al., 2006; Paulen and McClenaghan, 2014; Ross et al., 2019). There is no evidence of drumlins, eskers, or flutes occurring on Dundas Harbour. Indeed, landforms typical of glaciated landscapes, such as eskers, moraines and striations are rare on most of Devon Island (Dyke, 1999; Fortier et al., 1963; Grau Galofre et al., 2018). Glacial moraines are present along the southeastern coast of Devon Island (Dyke, 2001) in the proglacial environments of many piedmont glaciers feeding off of the Devon Ice Cap. Many of these moraines are visible in World Imagery (Esri, 2018) with evident push moraines present.

The landscape at the Dundas Harbour study site is peripheral to two fjords, one of which is still glaciated, and only a small area is covered by glacial till (Fig. 3.1). This is a much smaller-scale study site in comparison to currently documented ring-ridge moraines.
Additionally, Dundas Harbour is on an uninhabited, sparsely vegetated island that was likely very recently deglaciated based on its proximity to the Devon Ice Cap. Whereas currently documented ring-ridge moraines are large-scale, older in age (regions deglaciated ~10–14 ka BP), and heavily vegetated or farmed. We posit that vegetation, degradation, and farming may have removed any remnant small-scale features.

3.6. Origin of Dundas Harbour VRFs

We have identified circular, sinuous and anastomosing ridges within a till deposit in the Canadian High Arctic at Dundas Harbour, Devon Island, Nunavut, that we refer to as Vermicular Ridge Features (VRFs). These features share certain characteristics with periglacial landforms (Table 3.1), such as stone circles, sting bogs, and frost mounds, as well as glacial moraines (Table 3.2), such as ring-ridge moraines and kames. However, as discussed in the previous section, there are significant differences with previously reported landforms.

Based on our observations, we interpret Dundas Harbour VRFs to be primarily glacial in origin with minor post-depositional periglacial modification, such as thermokarst processes, differential upheave, thermal contraction cracking, and solifluction. We interpret these features to be akin to ring-ridge moraines as described by Knudsen et al. (2006) that form from the passive ablation of stagnant “dead” ice and subsequent deposition of supraglacial till (Fig. 3.12). However, there are some distinct differences from the circular ridges described by Knudsen et al. (2006) and other ring-ridge moraines that lead us to suggest the Dundas Harbour VRFs are a new/unique type of ring-ridge moraine. If these VRFs are ring-ridge moraine, then these are the first ring-ridge moraines identified in the Canadian High Arctic to the authors’ knowledge.

Ring-ridge moraines are a diverse glacial feature (Table 3.2), many of which can be found in the low and sub-Arctic zones of North America and Scandinavia (Corell et al., 2013) that deglaciated up to 14 ka BP. The limited examples of ring-ridge moraines that occur in clast-rich sandy till (e.g., Knudsen et al., 2006; Sutinen et al., 2014) are much larger in scale and associated with subglacial features that are absent at the Dundas
Harbour study site. We suggest that the difference in scale is a product of preservation, age, location, and documentation.

Figure 3.12. Vermicular Ridge Feature (VRF) formation model modified from circular ridge formation model by Knudsen et al. (2006). Over time (t) stagnated glacial ice under a layer of supraglacial diamicton will begin to down waste and separate into individual ice blocks (t₁–t₃). Supraglacial debris is transferred away from topographic highs and deposited into topographic lows via mass movements and meltwater action (t₁–t₂). Ice melts at unequal rates and deposits supra- and englacial debris around the edges of the melting ice blocks (t₁–t₃) and form adjacent ridges (t₃–t₄; outlined in black). Topography after all ice has fully ablated leaves a series of vermicular ridges and troughs (t₄). Closed cell ridges form where an ice block was once present, and a central concave trough is left behind. Present-day VRFs may be at any stage.

Human activity, farming practices, and vegetation may hide or destroy smaller scale features, such as those observed at Dundas Harbour, that may have once existed in the low and sub-Arctic continental moraines. Small-scale features may also be destroyed or periglacially reworked over time. Dundas Harbour VRFs are younger, sparsely vegetated, and undisturbed by human activity demonstrating a more pristine example of ring-ridge moraines. Yet, periglacial overprinting and reworking of material is already modifying
the VRFs at Dundas Harbour. Moraines along the coast of Devon Island are also much smaller in scale than the widespread continental moraines where ring-ridge moraines have been documented. This may result in smaller-scale landforms, and, therefore, smaller-scale ring-ridge moraines. Additionally, there are no documented ring-ridge moraines in the Canadian High Arctic. This absence may be the result of the rarity of this landform, the misidentification of patterned ground in the Canadian High Arctic, and/or the understudying of the Canadian High Arctic. There is also limited documentation of ring-ridge moraines composed of clast-rich till. This may suggest a limited database and/or a rare set of suitable environmental conditions to produce such landforms.

Ring-ridge moraine formation remains controversial. However, a clast-rich till with the grain size distribution curve observed in Dundas Harbour VRFs suggests low transport distances which would be expected in an ablation moraine. Additionally, Devon Island is largely absent of subglacial landforms, such as drumlins, eskers, and flutes, in general. These subglacial landforms are also not observed at the study site. Therefore, the Dundas Harbour VRFs are more likely to have formed from the passive ablation of stagnant “dead” ice and subsequent deposition of supra and englacial till rather than subglacial processes. This newly documented ring-ridge moraine may provide additional insight into this long-debated topic.

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Chapter 4: Vermicular Ridge Features Across Axel Heiberg Island, Nunavut

4.1. Introduction

The topographic inversion of glacial sediments due to the ablation of underlying glacial ice is a common mechanism for the production of hummocky surfaces (e.g., Clayton, 1964; Embleton and King, 1975; McKenzie, 1969). This process usually forms a series of mounds and depressions but has been observed to create conspicuous circular ice disintegration features (e.g., Gravenor and Kupsch, 1959) to sinuous and anastomosing ablation moraine ridges (e.g., Chapter 3; Knudsen et al., 2006). However, it can be difficult to differentiate between glacial landforms that form from this process based on morphology alone. Additionally, many periglacial features can produce morphologically similar features that can also form from differential ablation of underlying ice (e.g., Mackay, 1963, 1998) or in the absence of massive ice (e.g., Hallet, 2013). Sedimentology, depositional environment, and landform association are, therefore, important characteristics to consider when classifying a landforms (Embleton and King, 1975).

We have identified a landform across Umingmat Nunaat (Axel Heiberg Island), Nunavut that is morphologically, morphometrically and sedimentologically similar to Vermicular Ridge Features (VRFs) identified at Dundas Harbour on Tallurutit (Devon Island) in Chapter 3. We use VRF as a strictly descriptive term before assessing the mechanism of formation. A comparison to VRFs from Dundas Harbour and other morphologically similar features is used to assist in determining an origin. The VRFs observed on Dundas Harbour (Chapter 3) were morphometrically unique ablation moraines to the Canadian Archipelago. However, the widespread occurrence of very similar features across Axel Heiberg Island could suggest either this morphology/morphometry, mechanism or both is not uncommon across the Canadian High Arctic.
4.2. Geologic and Geomorphic Setting

Umingmat Nunaat, commonly referred to as Axel Heiberg Island, is located in the Qikiqtani region of Nunavut of Inuit Nunangat in Canada (Fig. 4.1). Axel Heiberg Island is also part of the Sverdrup Islands in the Queen Elizabeth Islands of the Canadian Arctic Archipelago.

Axel Heiberg Island has a complex geologic history. The island lies within the thickest section (up to 13 km) of the Sverdrup Basin which is predominantly composed of Carboniferous to Paleogene siliciclastics, evaporites, and carbonates (Balkwill, 1978; Embry and Beauchamp, 2008; Harrison and Jackson, 2014; Russell et al., 2006). The Sverdrup Basin was once a shallow sea that resulted from the opening of the proto-Arctic Ocean during the Early Jurassic to Early Cretaceous (Evenchick et al., 2019). This led to the deposition of a large amount of sands, silts and shale (Harrison and Jackson, 2014) which was interrupted by basaltic volcanism and the intrusion of diabase dykes and sills from the High Arctic Large Igneous Province (Evenchick et al., 2019). Deposition into the Sverdrup Basin eventually ended following the Eurekan Orogeny (Embry and Beauchamp, 2008) which uplifted and deformed basin strata and led to the intrusion of evaporites. Quaternary deposits, including stream, deltaic, glacial and marine beach sediments, overlie bedrock geology, valley floors and comprise raised beach sediments along the coasts (Thorsteinsson, 1971a, 1971b).

The island hosts two major ice caps, the Müller Ice Cap and Stacie Ice Cap (Fig. 4.1a), and a wide range of glacier types, such as cirque, outlet, piedmont, and valley glaciers (Omanney, 1969; Thomson et al., 2011). Axel Heiberg Island reached its last glacial maximum around 29 ka BP as part of the Inuitian Ice Sheet (Bednarski, 1998). Extensive deglaciation of the Inuitian Ice Sheet occurred predominantly from west to east between 16.5 and 11 ka BP and marine-based ice largely disappeared by 9 ka BP leaving mostly land-based ice on Axel Heiberg and other islands (England et al., 2006). Deglaciation of the island proceeded by freeing most of its fjords of ice by 8 ka BP (England et al., 2006) until reaching contemporary conditions. The marine limit varies
across the Axel Heiberg Island but ranges between 78 and 158 m (e.g., Bednarski, 1998; Dyke et al., 2005; Pollard and Bell, 1998).
Figure 4.1. Study site observed using World Imagery (Esri, 2018). (a) Umingmat Nunaat, also known as Axel Heiberg Island, is located in Nunavut, Canada. Nansen Sound runs along the east coast of much of Axel Heiberg Island. White box locate panel b. MIC and SIC represent Müller Ice Cap and Stacie Ice Cap, respectively. (b) Location of field site (located in white box) is northeast of Mokka Fiord which feeds into Eureka Sound. World Imagery Source: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community.

The main field area is located ~ 20 km east of Mokka Fiord and ~ 12 km southwest of Mokka Fiord Diapir (Fig. 4.1b). The field site lies within the Granite dispersal train sourced from the Precambrian Shield located in southeastern Ellesmere Island (England et al., 2006; Ó Cofaigh et al., 2000) and is composed of Quaternary deposits (Thorsteinsson, 1971a, 1971b). The region where we identified and ground-truthed VRFs in the field is located on a terrace adjacent to a channel trending northwest-southeast feeding glacial meltwater from the highlands into Mokka Fiord. Although the field site is not directly analysed for surficial geology and geomorphology in previous studies, approximately 300 km to the northwest along the coast of Nansen Sound and Flat Sound, the landscape is dominated by meltwater channels sourced from the western highlands, moraines and kame terraces, and marine sediments (Bednarski, 1998).

Present-day conditions represent a polar desert environment (Andersen et al., 2002). The nearest long-term climate station is Eureka A station located on the coast of Fosheim Peninsula on Mirnguiqsirvik (Ellsemere Island) ~ 300 km northeast of the field site. The Eureka A station reports a mean annual air temperature of −18.8°C and a mean annual precipitation of 79.1 mm (mostly in the form of snow—60.3 mm) between 1981 and 2010 (Environment Canada, 2021). Permafrost thickness has been measured to be 400 to 600 m at Mokka Fiord (Pollard et al., 1999; Taylor and Judge, 1976).
4.3. Methodology

This study involved fieldwork in July 2019 at Mokka Fiord, Axel Heiberg Island, followed by satellite and aerial imagery mapping across Axel Heiberg Island.

4.3.1. Field Region

Fieldwork was carried out in July 2019 21 km east of Mokka Fiord on Axel Heiberg Island. Field reconnaissance via hiking and helicopter led to the identification of VRFs across multiple terraces along the same channel. One terrace was selected for in-depth field analysis, including trenching, Light Detection and Ranging (LiDAR), and Ground Penetrating Radar (GPR) data collection to characterize the landforms.

4.3.1.1. LiDAR

An AkhkaR4DW backpack mobile laser scanning system was used to kinematically collect high-precision 3D topographic data (Hyyppä et al., 2020; Kukko et al., 2017, 2012, 2020; Liang et al., 2015). This Light Detection and Ranging (LiDAR) system was developed by the Finnish Geospatial Institute to produce ultra-high resolution (1‒5 cm-scale) digital elevation models (DEMs). The positioning system relies on GPS and GLONASS constellations and operation is based on Global Navigation Satellite System – Inertial Measurement Unit (GNSS-IMU).

Two precision profiling laser scanners collect 3D data at different wavelengths, including the primary Riegl VUX-1HA (Riegl Gmbh, Austria) and the secondary Riegl miniVUX-1UAV scanners. The primary scanner operates at 1550 nm wavelength and collects high density data using 1017kHz pulse repetition with a scanning rate of 250 lines per second. This provides a 3D location for each measured point and information of the surface reflectivity and signal amplitude. The secondary scanner operates at 905 nm and has a pulse repetition rate of 100 kHz with a scanning rate of 100 lines per second. The primary scanner is positioned in the vertical scan plane position and the secondary scanner is tilted 25 degrees, both of which provide a 360 Field of View (FoV). Mapping using the LiDAR system is carried out by Dr. Antero Kukko at walking speeds around 3.5 km/h.
depending on the terrain which provides an along-track scan line spacing of 3.3–5.5 mm for VUX-1HA, and 8.3–13.8 mm for miniVUX-1UAV. A maximum range of 125 m on dark objects can be reached. However, highly reflective objects can be captured at longer ranges.

WhiteBox Geospatial Analysis ToolBox (GAT) is an open-source geospatial data analysis software developed by Professor John Lindsay at the University of Guelph (Lindsay, 2014, 2016). This software provides comprehensive functionality for LiDAR data analysis. The LAS file, a standard binary file format for the interchange of lidar data, was used to create a Bare Earth DEM which was used to create a Hillshade file. The LAS file was produced by Dr. Antero Kukko using TerraScan software and had a total of 46,163,219 points with an average density of 164.2 points/m².

The Bare Earth DEM was created using an inverse-distance weighting (IDW) scheme. IDW interpolation assumes points close to each other are more alike that those that are farther. IDW uses measured point values to predict unmeasured point values. A search distance of 10 cm was used to interpolate the point cloud. Measured point values closest to the 10 cm search distance (diameter) are weighted more heavily (have more influence) on the predicted values than those farther away. Therefore, the weight of a point value is a function of the inverse of the distance raised to the power of p. The rate at which the weights of the point values decreases is dependent on the Power (p) exponent. The default value of 2 is used to process our lidar data which will give more weight to nearby points and less weight to points farther away which will ultimately provide more detail in the DEM. Points that exceed a slope of 30° from the unmeasured point being calculated are considered an outlier/non-ground point and were not used in the output point-cloud. A grid resolution of 5 cm/pixel was used to provide a high-resolution DEM with lower processing time.

A hillshade map was created in WhiteboxGAT by calculating an illumination value for each cell in the Bare Earth DEM. The azimuth (direction of the sun), measured clockwise from North, was set to 315° (northwest). The altitude (angle of illumination), measured
from the horizon to normal, was set to 30°. The Bare Earth DEM and Hillshade files were loaded into ArcGIS Desktop 10.8.1 using a WGS 1984 UTM 16N projection.

4.3.1.2. GPR

Sensors and Softwares’ 250 NOGGIN SmartTow GPR system was used to search for massive ice and deposit thickness with 250 MHz antennae. Five GPR lines were collected, three of which lie within the lidar data, ranging from 20 to 30 meters in length. Signal velocity was determined based on sedimentology, diffraction hyperbola fitting, and context from trenching in the field which was determined to be 0.125 m/ns (frozen and unfrozen sand and gravel). Based on this signal velocity, GPR signals penetrated down to roughly 4 m before heavily attenuating. GPR data was collected on July 8, 2019 and therefore the thaw depth is representative of that day of the year.

GPR data is analyzed using Sensors and Softwares’ Ekko_Project_5 software. GPR lines were dewowed (i.e., signal saturation correction) which removes unwanted low frequency signals (known as the ‘wow’) that is usually the product of the proximity of the transmitter and receiver, as well as the electrical properties of the ground. Dewow applies a running average filter on each trace within a window with a width equal to the pulsewidth at the nominal frequency of that trace. The average of point values within this window is subtracted from the central point and continued at each point along the trace.

The GPR signal strength was amplified with a Spherical Exponential Calibrated Compensation (SEC2) gain to enhance weaker signals that result from spherical spreading losses and exponential ohmic dissipation of energy. However, signal strength is strongest at the surface and reduces with depth. The SEC2 gain compensates for this by having a lower Start Gain value (4.5 used here) that increases with depth until the specified Maximum Gain (1000 used here) is reached. The Attenuation value (8 used here controls the exponential rise (slope) from the Start Gain to the Maximum Gain and has the biggest effect on how the data is displayed.

Elevation data along each GPR line is extracted from the lidar dataset and added to the GPR data file. To do this, the GPR lines are first georeferenced to the high-resolution
lidar dataset, which has higher location accuracy than the built-in GPR GPS. This is done manually by comparing lidar data to field images and GPS points to identify the accurate start and end points along the GPS line collected. The elevation data along this line is then input into Ekko_Project_5 and the cross-section is shifted to compensate for topography. This corrects for unreliable depths of key subsurface features, but slightly stretches the upper part of the cross-section image.

4.3.2. Mapping

Mapping of Vermicular Ridge Features across Axel Heiberg Island was pursued to better understand the depositional environment to consequently provide context into formational processes involved. It is important to note that landforms identified as VRFs have not been ground-truthed aside from the study site and locations where VRFs were observed and photographed nearby from a helicopter. These examples from the field are used as the basis for assisting in the identification of similar features across the island.

World Imagery (Esri, 2018) accessed from the online ArcGIS database was used to identify and map landforms across Axel Heiberg Island in ArcGIS Desktop 10.8.1 using a WGS 1984 UTM 16N projection. Grid-mapping was carried out with an average grid size of 4,185 km² to map the entire island. We mapped approximately 80% of the island. Regions with ice, snow, or cloud cover, as well as poor lighting conditions, were mapped as indeterminate regions. Examples of landforms that looked most similar to VRFs characterized in the field were mapped as VRFs, whereas landforms that appeared similar but varied in morphology from VRFs observed in the field were mapped as questionable. The Northwest and southeast regions of the island have yet to be mapped.

4.4. Results

4.4.1. Field Region

We identified multiple terraces along the same channel in the field that hosted VRFs (Fig. 4.2). VRFs occur in a homogenous deposit composed of subrounded to rounded silt, sand, pebbles, and cobbles. A thin white salt crust can be found over much of the deposit.
The deposit is cut by the channel exposing a ~6 m thick talus slope that transitions into a ~12 m thick gentler talus slope characterized by lobes before connecting with the riverbed (Figs 4.2 and 4.3a) suggesting the deposit has a maximum thickness of 18 m at the river cutbank. An 89 cm deep pit was dug into the deposit without reaching the thaw depth (July 2019) (Fig. 4.3b). There is minor evidence for a preferred flat orientation of large grains (Fig. 4.3b). Less cobbles were present below 70 cm. The deposit appears to thin northward and thicken southward locally at the field site. Additional VRF terraces were identified north and south of the field region along the same channel via helicopter reconnaissance (Figs. 4.2 and 4.4).

The main field terrace has an average elevation of 143 m (Fig. 4.2). Six additional terraces host VRFs based on field reconnaissance (Fig. 4.2). Five terraces, including the main field terrace, reside on the western side of the channel and two terraces reside on the eastern side (Fig. 4.2). The highest terrace is on the western side with an average elevation of 166 m. Three other terraces on the western side are lower than the terrace where the field site is location, with elevations of 131, 130, and 126 m. The two eastern terraces have elevations of 129 and 114 m. The uppermost terrace (at 166 m) had less pronounced, and possibly more degraded, VRFs.

VRFs found at the field region and from the helicopter exhibit a circular, elongate, sinuous and/or anastomosing ridge and trough morphology in planform (Fig. 4.4). Raised convex ridges stand above the rest of the deposit and frequently encircle a central concave depression (Figs. 4.4–4.6). Ridges can exist as circular (Figs. 4.4b, c), semi-circular, elongated, or sinuous closed cells (Figs. 4.4a, c, e, f). The central depressions sit at the same elevation or higher as the mesh with the ridges elevated above its adjacent terrain (Figs. 4.5, 4.6). Ridges can also be subdued, shallow and wide compared to the more prominent narrow convex ridge (Fig. 4.5). Small sharp-crested conical mounds can be found in the same deposit as VRFs (Fig. 4.4d).

Ridges can reach up to 1.5 m in height when measuring from the ridge apex to the adjacent low-lying terrain. However, ridges often do not exceed 1 m in height. Closed-cell ridges range in height between 0.2 and 0.6 m when measuring from the lowest point
in the central trough to the highest point on the ridge (Fig. 4.6). Ridge width ranges between 1.5 and 9 m but more commonly ranges between 3 and 4 m from ridge to ridge. Thirty-two closed-cell ridges with central troughs were identified in the LiDAR area (Fig. 4.5). The long axis of closed-cell ridges ranges between 5.8 and 36.8 m with an average of 15.8 m. The orientation of the long axes (North = 0°, East = 90°, South = 180°) range between 1.8° and 174.5° with an average of 95.7° and therefore have an average East-West orientation. The short axis of closed-cell ridges ranges between 4.3 and 12 m with an average of 8.2 m.
Figure 4.2. Field region of VRF terraces. Southern part of Mokka Fiord Diapir labeled. Seven terraces with VRFs were ground-truthed (outlined in white). The elevation (in meters) of each terrace is numbered in white. The elevation with an astrisk denotes the main field terrace for in-depth field analysis (i.e., GPR, LiDAR, and trenching). Black arrow points to location of riverbank in Figure 4.3a. White points indicate additional VRF deposits identified via helicopter (not ground-truthed) and figure 4.4 locations.

Figure 4.3. VRF deposit characteristics. (a) River cutbank exposing deposit thickness (Location identified in Figure 4.2). Polygons are visible at the top of the deposit at this location. The top of the deposit to the white dashed line is characterized by a steep talus slope ~ 6 m thick. A lowering of slope occurs between the white dashed line and the white dotted line which is characterized by lobes of talus. Below the white dotted line is a river sand bank. (b) Pit dug 89 cm into main field terrace. White dashed line outlines flat-oriented gravel. Red dotted line marks 70 cm where a change to more sandy material occurs with less clasts.
Figure 4.4. Examples of VRFs in the field as seen by helicopter. Figure locations can be found in Figure 4.2. (a) VRFs at the main field terrace near Mokka Fiord looking North.
Mokka Fiord Diapir in the foreground to the East. Cracks can be seen along or just off of the axial trace of some of the coarse-grained ridges. (b) VRFs on terrace on the opposite side of the channel in the field region. (c) North of field region, directly west of Mokka Fiord Diapir. Polygonal troughs cross-cut VRFs. (d) Sharp-crested mounds and VRFs south of field site. (e) VRFs north of field region, west of Mokka Fiord Diapir. (f) Linear VRFs in dark-toned deposit directly west of Mokka Fiord Diapir. Polygonal troughs cross-cut VRFs. (g) VRFs near Strand Fiord. Polygonal troughs cross-cut VRFs.

Figure 4.5. DEM and Hillshade of field site VRFs. DEM is overlying hillshade with 315° azimuth. GPR transects 1–3 outlined in red. Topographic profiles and GPR transects can be found in Figure 4.6. Features of note include, (a) shallow and wide raised ridges, (b) perfectly circular VRF, (c) polygonal trough cross-cutting VRF, and (d) raised polygon shoulder.
Figure 4.6. Topographic profiles and GPR transects of VRFs. Transect locations can be found in Figure 4.5. Units in meters. Elevation is on the y-axis of GPR transects. Line 1 (A-A’), Line 2 (B-B’) and Line 3 (C-C’) topographic profiles demonstrate pointed to rounded convex VRF ridges surrounded by flat-floored mesh. VRF troughs are pointed to bowl-shaped concave depressions. GPR transects indicate a thaw depth (i.e., depth to permafrost in July 2019) (yellow dashed line) occurring 1–1.5 m below the surface. Bright reflectors can be seen beneath closed cell VRF central troughs indicating the presence of ice. These reflectors demonstrate a wedge-shaped object in Lines 1 and 2 (blue dotted line).

Vegetation acts as a distinguishing factor between the ridges and its surrounding (Fig. 4.4). Central depressions and much of the topographic lows throughout the deposit host vegetation, such as grasses and mosses, whereas the ridges only host lichens (black, orange and white).

A crack can be found running along or just off-centre from the axial trace of many ridges (Fig. 4.4a). Closed-cell ridges generally have a crack running off-centre from the axial travel along the inner part of the cell. Cracks present themselves as a thin and narrow cavity along the ridge where slumping of the surrounding material has occurred. Cracking also occurs along the center of polygon troughs and along the shoulders of polygons. These cracks tend to be much wider (≤ 30 cm) and deeper.

Polygons range in diameter (long axis) between 115 and 167 m. Polygon trough width averages around 3 m but can reach up to nearly 6 m. Thin and narrow secondary troughs are present within the larger polygon centers (Fig. 4.5). Secondary troughs propagate from a major trough and often terminate within the polygon center. Polygon troughs appear to cross-cut VRFs but can also merge with VRFs to create an anastomosing ensemble of ridges and troughs (Figs. 4.4 and 4.5).
Figure 4.7. Thaw slumps exposing VRF deposit thickness. (a) West side of channel, south of field site. Possible massive ice exposed at thaw sump overlaid by VRF deposit. Thaw slump exposure is around 10−15 m with ~1−2 m of material above. (b) East side of channel, directly opposite of field site. VRF deposit overlies a lighter-toned deposit. Darker-toned deposit thickness is around 10−15 m. (c) West side of channel, south of field site. VRF deposit overlying lighter-toned sediments. Deposit thickness roughly 10−15 m.
Five GPR transects were collected at the main field site and show a thaw depth ranging between 1 and 1.5 m that can be identified by a nearly continuous bright linear reflection below the surface. Lines 1, 2, and 3 cross through closed-cell ridges (Figs. 4.5 and 4.6). A bright radar reflection can be identified at the frost table below the central depression of a close-cell ridge at Lines 1 and 2. This bright reflection indicates ice is present. Below this bright reflection an ice-wedge can be resolved (Fig. 4.6). Line 3 shows bright reflections indicative of ice beneath two troughs adjacent to a closed-cell ridge. However, the center of the closed-cell ridge does not appear to have an obvious bright reflection indicating subsurface massive ice. Three separate polygon troughs were observed with the GPR. Two primary polygon troughs show bright reflections at the frost table indicating the presence of an ice wedge. However, the structure of the ice wedge cannot be resolved. One secondary polygon trough was intersected by the GPR and no obvious bright reflection can be identified, indicating an ice wedge is either not present or too small to resolve. Deposit thickness is indeterminate in the GPR transects, suggesting the deposit thickness exceeds the signal penetration depth of approximately 4 m.

Active thaw slumps were observed by helicopter on many of the terraces cutting into the VRF deposits (Fig. 4.7). VRFs were observed directly on top of thaw slumps. Thaw slumps consistently exposed an approximately 10 to 15 m thick deposit.

### 4.4.2. Mapping

Features reminiscent of VRFs observed in the field were mapped across Axel Heiberg Island based on planform morphology (Figs. 4.8 and 4.9). Features that resembled VRFs but appeared subdued or morphologically different from planform VRFs that were ground-truthed at the field site and by helicopter were categorized separately as possible VRFs. An area of 14,150 km$^2$ (32% of the island) was considered to be indeterminate due to poor lighting or resolution, or glacial, snow or cloud coverage. VRFs covered a total of 29.2 km$^2$ of the island, which includes 15.3 km$^2$ being VRFs and 13.9 km$^2$ being possible VRFs.
Figure 4.8. VRF mapped across Axel Heiberg Island using World Imagery (Esri, 2018). (a) VRFs (red) and possible VRFs (green) mapped across Axel Heiberg Island. White box indicates panel b. (b) VRFs mapped along the eastern coast of Axel Heiberg Island. A two-toned contact can be observed running along the eastern coast. White box indicates location of panel c. (c) Field and helicopter region near Mokka Fiord mapped. A two-toned boundary can be observed running north-south. World Imagery Source: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community.

Morphologic variation exists among VRFs mapped on Axel Heiberg Island (Fig. 4.9). VRFs can be isolated or can anastomose into a brain-like form (Fig. 4.9). A range of circular, elongated, sinuous and anastomosing VRFs occur. Vegetation, thermokarst and two-toned deposits can accentuate the VRF troughs from ridges. Possible VRFs have a similar morphology, but can appear subdued, inverted, or elongated on a slope and appear as if other formation and/or alteration mechanisms may be at play. Polygons of varying shapes and sizes are often present where VRFs occur. Polygon troughs appear to cross-cut closed-cell VRFs and/or share ridges along the polygon shoulders (Fig. 4.9).

VRFs are more common along the eastern coast of the island, most of which occur above the marine limit. However, some examples are found below the marine limit close to the
coast. The eastern side of the island is composed predominantly of gently sloping meltwater channels, whereas the western side of the island is mountainous. VRFs are consistently found alongside channels draining from the highlands. This can create a string of VRFs occurring alongside a channel across a few kilometers. VRFs are not found on steep slopes or in mountainous regions. Ice marginal examples of VRFs are rare which are characterized as possible VRFs.

Figure 4.9. VRFs mapped across Axel Heiberg Island using World Imagery (Esri, 2018). (a) -87.5455, 79.6291. (b) -87.7371, 79.6602. (c) -89.6661, 80.3875. (d) -87.5585, 79.6846. (e) -87.5908, 79.7782. (f) -95.7332, 79.7364. (g) -90.6567, 80.494. (h) -85.9819, 79.4053. (i) -87.7078, 79.8002. World Imagery Source: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community.
4.5. Discussion

A landform referred to here as Vermicular Ridge Features (VRFs), as a purely morphologic descriptor, has been identified northwest of Mokka Fiord on Axel Heiberg Island, Nunavut, Canada in the field, and subsequently mapped across the island. VRFs, forming a circular, elongated, sinuous or anastomosing morphology, can be found across Axel Heiberg Island, particularly on the eastern side of the island where highlands glaciers drain meltwater and sediment across gently sloping plains into Nansen Sound, Flat Sound, and Eureka Sound.

VRFs identified at Dundas Harbour on Devon Island, Nunavut (Chapter 3) have an analogous morphology to VRFs identified at Axel Heiberg Island. VRFs at Dundas Harbour are proposed to be the result of passive ablation of stagnant “dead” ice and the subsequent deposition of supra- and englacial debris, as has been proposed for other morphologically similar circular rimmed mounds (i.e., ring-ridge moraines) in the glacial community (e.g., Aartolahti, 1974; Ebert and Kleman, 2004; Gravenor and Kupsch, 1959; Knudsen et al., 2006; Ross et al., 2019). Below, we compare observations made at Axel Heiberg Island to VRFs at Dundas Harbour and other morphologically similar glacial landforms in order to elucidate an origin.

Ridges at Dundas Harbour were dubbed VRFs due to their planform circular, elongate, sinuous and anastomosing (or brain-like) morphology. VRFs at Dundas Harbour share many characteristics to VRFs at Mokka Fiord on Axel Heiberg Island, particularly VRF morphometrics (i.e., ridge height, width, closed-cell diameter). VRF ridges at Dundas Harbour exhibit heights between 0.5 to 2.5 m (more commonly 0.5 to 1 m) and widths between 1.5 and 6 m, whereas VRF ridges at Mokka Fiord exhibit heights between 0.2 and 1 m and widths between 1.5 and 9 m (more commonly 3 and 4 m).

Additionally, topographic profiles of circular examples of VRFs on Dundas Harbour and Mokka Fiord have similar micromorphology with rounded convex ridges and a central concave u-shaped central depression (Fig. 4.10). VRFs on Dundas and at Mokka also exhibit a crack running along or just off of the axial trace of ridges, separate from the
polygonal thermal contraction cracks thaw cross-cut VRFs at both sites. Polygons, however, are much more developed at the Mokka Fiord field site compared to Dundas Harbour. This is likely the product of longer subaerial exposure at Mokka Fiord as the field site is much farther away from the Ice Caps compared to the VRFs located at Dundas Harbour.

Figure 4.10. Topographic profiles of circular closed cell VRFs at Mokka Fiord and Dundas Harbour. Example of circular VRF at Mokka Fiord (D-D’) is ~half the size of example of circular VRF at Dundas Harbour (E-E’) from Chapter 3. Microtopography of both topographic profiles should gradual outward-facing slopes, rounded convex ridges, and u-shaped central troughs. The abrupt change in slope from ridge to central trough is caused by a crack along the axial trace of the ridge.

The sedimentology of VRFs at Dundas reflect that of a sandy clast-rich till that experienced low transport distances and had no clear evidence of stratification, which eliminated a glaciofluvial origin (i.e., kames and kettles). Although no grain size distribution was done on the VRF deposit at Mokka Fiord, the material is also clast-rich sand, but clasts are more rounded and show some evidence of preferred orientation and
stratification of sands and gravels. This sedimentology is more indicative of a glaciofluvial environment.

Ridges of varying shapes and sizes are common features produced by glacial processes. However, ice stagnation processes most commonly create circular to chaotic mounds, ridges, and depressions similar to VRFs observed at Axel Heiberg Island. Supraglacial hummocky moraines, controlled moraines, kame and kettle topography, ice-walled lake plains, kame terraces, and pitted sandar (i.e., pitted outwash plain/terrace) all form a series of ridges and troughs due to the ablation of stagnant ice and deposition of glacial sediment. Due to similar formation mechanisms and depositional environments, these features are often associated (Benn and Evans, 2010).

Kames are mounds of bedded sands and gravels formed on top of or alongside a glacier via glaciofluvial sediment deposition. Kettles are depressions of subsidence following the melting of buried ice or deposition of glaciofluvial sediment around an ablating isolated ice block. Therefore, kame and kettle topography comprise mounds, ridges and depressions, some of which are circular. Kame and kettle topography can easily be confused with hummocky moraines; however, these can be distinguished by sedimentology and the presence of remnant glaciofluvial terraces (Benn and Evans, 2010).

Bednarski (1998) describes the Quaternary geomorphology and stratigraphy of northeastern Axel Heiberg Island, which includes our mapping area and is ~300 km northeast of our field site, and suggests the presence of kame terraces and kettled outwash terraces. They note the sudden change of tone and elevation that distinguishes the outwash plains emanating from the Princess Margaret Range from the lighter-toned marine sediments along the lower slopes of the eastern coast and mark this as the marine limit (~78–158 m).

The outwash plains described by Bednarski (1998) are suggested to host extensive kettled outwash terraces composed of coarse gravel that end at the marine limit, where the lowlands along the coast can have outwash material that is overlain by marine sediments and dropstones (Bednarski, 1998). These terraces are suggested to be ice-contact
glaciofluvial systems composed of glaciofluvial gravel with kettles and kames ranging from 30–50 m in relief, ice-contact ridges, and active slumping that indicates the presence of buried glacial ice. Kame terraces are also suggested to be present along the western coast of Flat Sound where a northwest-flowing trunk glacier in Nansen Sound was in contact with the eastward-flowing glaciers from the Axel highlands (Bednarski, 1998).

Kame terraces are ice-marginal/ice-contact features that form alongside meltwater channels and were in contact with glacial ice and can be easily confused with fluvial or outwash terraces (Menzies, 2002). However, kame terraces will typically have multiple steps/terraces on one side of a channel and will have varying gradients and elevations to terraces on the opposite side of the channel (e.g., Gray, 1975; Sissons, 1982). Kame terraces are mostly composed of glaciofluvial sands and gravels from lateral meltwater channels; however, supraglacial and englacial till can accumulate on the surface (e.g., Levson and Rutter, 1989). Kame terraces are often associated with kettle and kame topography, hummocky moraines, and eskers (Benn and Evans, 2010). Finding other remnant glaciofluvial terraces can be one simple way of distinguishing kames and kame terraces from hummocky moraines. Kames and kame terraces are composed of glaciofluvial materials and are generally stratified as a result. Ice blocks can become stranded in a kame terrace or outwash plain which can act as an obstacle for glaciofluvial sediment to accumulate around or can become completely buried (e.g., Russell et al., 2006).

The observations made by Bednarski (1998) are consistent with observations at Mokka Fiord and mapping the eastern coast of Axel Heiberg Island. A two-toned contact (Fig. 4.8) with a sudden change in elevation continues southward into our field site, with the exception of the Mokka Fiord Diapir and Gibs Fiord Diapir that are elevated above much of the area (> 600 m asl). Therefore, we consider the field site to be located near the marine limit. We also observed VRFs in the coastal lowlands which are light-toned (Fig. 4.8) and could represent VRFs overlain by a veneer of marine sediments.
The VRFs at Mokka Fiord reside on multiple terraces along a meltwater channel trending NW-SE, which is the same orientation as the proposed kame terraces near Flat Sound (Fig. 4.8). The presence of multiple terraces with uneven elevations, sedimentologic observations, a meltwater channel orientation reflecting the contact of the proposed trunk glacier in Nansen Sound and eastward-flowing glaciers from Axel Heiberg (Fig. 4.1), and the presence of mounds (i.e., Fig. 4.4e), ridges, and troughs, suggests the VRFs at Mokka Fiord are part of remnant kame terraces. Mounds (i.e., Fig. 4.4e) are interpreted to be kames that, if ice-cored, may degrade into VRFs or remain kame mounds.

Thaw slumps (Fig. 4.7) and active lobate slumping (Fig. 4.3) suggest the presence of buried massive ice, likely to be remnant glacial ice, that underlies the VRF deposit. It is common for kame terraces to bury and preserve remnant glacial ice (e.g., McKenzie, 1969; McKenzie and Goodwin, 1987). As the ice melts, the deposits overlying it redistribute to preferentially promote a hummocky surface, much like the process described in Chapter 3. Closed-cell VRFs are almost certainly the by-product of differential ablation (e.g., Maizels, 1992; Chapter 3), and the subsequent collapse of the material may be what produces the “cracks” observed along the axial trace of the ridges (Fig. 4.4a). These “cracks” could represent subsidence which can create normal “gravity” faults from collapse (e.g., Branney and Gilbert, 1995; McDonald and Shilts, 1975) or high-angle convex upwards reserve faults (e.g., McDonald and Shilts, 1975). However, these cracks may be a by-product of periglacial modification via differential frost-heave as suggested in Chapter 3.

The kame terraces at Mokka Fiord have been exposed to subaerial conditions for ≤ 10 ka (Bednarski, 1998; England, 2006) making them prone to periglacial and paraglacial modification. Thermal contraction crack polygons are prominent across the kame terraces, and clearly cross-cut the VRFs indicating post-depositional modification of the VRFs. Some polygon shoulders are raised and composed of the same coarse-grained material at the VRFs. Distinct polygon shoulders with steep-sided troughs, as is seen at the Mokka Fiord field site, suggests actively growing wedges in the polygon trough (Hallet, et al., 2011). As thermal contraction cracks open in the winter, the crack becomes filled with water, snow, sand or dust during the warmer months. As this crack reopens
each year, new material is deposited into it and a wedge begins to develop. This creates horizontal compressive stresses that cause uplift along the shoulders of the polygons that will continue to grow as the wedge grows (e.g., Hallet et al., 2011). If the thermal contraction crack were to cross a VRF, then the resulting polygon shoulders would be raised higher than the average ridge height, which is observed at Mokka Fiord (Figs. 4.4 and 4.5). Cracks along the polygon shoulders would likely be the result of extensional stresses at the apex of the ridge and slumping along the steep-sided trough walls. The central depressions of closed-cell VRFs show indication of ice (Fig. 4.6). This is likely the result of a positive feedback loop where water collects and saturates the ground at the central depression. With the cold atmospheric temperatures and a saturated ground, frost cracks can open (MaKay, 1986) within the central depression, and possibly propagate outwards to cross-cut many of the VRFs. Therefore, polygonal cracks, raised polygon shoulders, and ice wedges are all considered to be periglacial modification of the VRF deposits post deposition.

Although the VRFs at Mokka Fiord are interpreted as kame terraces, VRFs are found across the island and may have varying origin. Bednarski (1998) suggested much of the eastern coastal plains to be kettled outwash terraces (i.e., pitted sandar). Pitted sandar form similarly to kame terraces, as they are another ice-marginal landform composed of glaciofluvial sediment and with its surface topography heavily influenced by differential ablation of buried or stranded glacial ice (e.g., Thomas et al., 1985). However, sandar can develop in a supraglacial, lateral marginal, or proglacial environment do not always occur as terraces (Benn and Evans, 2010). Therefore, it is likely that VRFs across the island are genetically similar being that they are the result of ice-contact/ice-marginal depositional settings and range from supraglacial, ice-marginal and pro-glacial environments.

4.6. Conclusions

A landform descriptively termed Vermicular Ridge Features (VRFs) are identified on Axel Heiberg Island and mapped across the island. VRFs have a circular, elongated, sinuous and/or anastomosing morphology that have similarly been identified at Dundas Harbour on Devon Island (Chapter 3). VRFs investigated near Mokka Fiord are
suggested to be hosted on remnant kame terraces, potentially from the separation of the eastward-flowing ice from Axel Heiberg Island and the northwestern-flowing ice of the trunk glacier occupying the Nansen Sound, as was proposed for the kame terraces near Flat Sound north of our field site (Bednarski, 1998). Remnant glacial ice is possibly the source of many thaw slumps underlying the VRF kame terraces in the area. VRFs across Axel Heiberg Island are mapped based on planform morphology. VRFs likely form as a result of paraglacial differential ablation of buried ice from glacial margins or detached ice blocks and periglacial modification. This deposit would represent either a supraglacial or ice-marginal depositional environment. Active glacial karst/disintegration used to be common in the late Wisconsin when widespread on stagnant glaciers covered the northern Great Plains of North America (e.g., Clayton, 1964; Gravenor and Kupsch, 1959). The widespread occurrence of VRFs on Axel Heiberg Island indicates that active glacial karst and ablation of buried ice is still common across the Canadian High Arctic today and may represent “fresher” versions of similar relict Wisconsinian ablation features observed across North America.

4.7. References


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Chapter 5: Discussion and Conclusions

Orbital parameters, such as eccentricity and obliquity, play an important role in shaping a planet’s climate. Mars has experienced a wide range in obliquity in its past (Laskar et al., 2004), affecting the distribution of insolation and, thus, ice deposition and stability (e.g., Madeleine et al., 2009). Presently, Mars has a similar obliquity (25°) to Earth (24°) which leads to increased insolation in the equatorial latitudes, and ice stability near the poles. Subsurface ice has been detected (e.g., Boynton et al., 2002; Feldman et al., 2002; Morgan et al., 2021; Pathare et al., 2018) across the mid-latitudes of Mars (e.g., Figs. 1.3, 1.4, and 1.6) and is suggested to be remnants from past obliquity variations when ice was more stable at the equatorial and mid-latitudes. However, the global distribution and abundance of subsurface ice is not well constrained as Shallow Radar (SHARAD) measurements are unable to resolve the upper ~5–10 m (e.g., Seu et al., 2007; Bramson et al., 2015) of the subsurface and other instruments used to detect subsurface ice, such as the Gamma Ray Spectrometer (GRS) Neutron Spectrometer (NS) cannot resolve deeper than 1 m (Pathare et al., 2018). Thus, it is unclear at what depth pure ice is present 1–10 m below the surface (Bramson et al., 2021). Depth to the top of these shallow ice deposits is extremely important when estimating ice volumes, determining past climate and present atmosphere-subsurface interactions, and assessing resource accessibility for future in situ resource utilization (ISRU) missions and human exploration (Abbud-Madrid et al., 2016). Therefore, as we reach the limitations of current datasets, the integration of multiple datasets and techniques becomes critical to search for subsurface ice on Mars.

Human exploration, and eventual habitation, of Mars represent one of the most exciting and challenging frontiers of human endeavours over forthcoming decades. Target landing sites for the SpaceX Starship mission to Mars (Fig. 1.1) have recently narrowed their focus to Arcadia Planitia, Phlegra Montes and Elysium Montes in Mars’ northern mid-latitudes (Golombek et al., 2021). These areas have been selected because they show indication of a significant supply of water ice surviving from previous Martian ice ages. However, the accessibility (i.e., depth to the ice), ice thickness, and ice purity at each site
is unknown. This dissertation aims to assist in constraining the location and abundance of near-surface water-ice at the proposed human mission landing site, Arcadia Planitia.

Prior to the recent down-selection of landing sites (Golombek et al. 2021), we targeted Arcadia Planitia due to evidence for significant subsurface ice at low mid-latitudes (e.g., Bramson et al., 2015; Pathare et al., 2018), and its low-lying and flat nature ideal for the safe decent, entry and landing of a spacecraft (e.g., Fig. 1.2). In order to further assess Arcadia Planitia as a potentially favourable landing site for ISRU and human exploration, we characterized site-specific ice-related landforms (Chapter 2, Appendix 1) in order to reconstruct the glaciological history (Chapter 2) and present conditions (Appendix 1) of the landing region. This will help in identifying the location and properties of ground ice for ISRU and climate studies, as well as the scientific value of the proposed landing region. We utilized Earth analogues to assist in the identification of relationships between the surface morphology of ice-related landforms and the subsurface ice content and depth to ice (Chapter 3, Chapter 4).

Global-scale maps utilizing multiple orbital datasets (e.g., thermal, neutron, radar, imagery) identified that Arcadia Planitia has a high probability of hosting significant ice (Fig. 1.3; Morgan et al., 2021). In Chapters 2 and 3 we utilized many of these same datasets (Table 2.1), in addition to HiRISE imagery, to identify small-scale site-specific variations in thermal, radar and morphologic data indicating the presence of near-surface ice. Although a buried ice sheet has been proposed to exist in Arcadia Planitia (Bramson et al., 2015, 2017) (which includes the study site from this dissertation), six prominent mapped units interpreted to be buried massive glacial ice remnants were identified (Fig. 2.2). These six glacial-related features consist of lobate debris aprons, ice mantling units, and features analogous to terrestrial ice streams (Fig. 2.12). These glacial features have distinct differences in TES and OMEGA albedo (Fig. 2.3), relative brightness in HiRISE and CTX (Fig. 2.5), morphology (Figs. 2.5, 2.6, and 2.10; Appendix 2), and thermal properties (Figs. 2.3, 2.6, and 2.7) from the surrounding terrain indicating a difference in surface and near-surface properties (Table 2.2). For example, the glacial features mapped in Chapter 2 exhibit moderate thermal inertia and low TES dust and albedo (Table 2.2), which we interpret to represent a low thermal inertia regolith overlying a high thermal
inertia ground ice. This interpretation is supported by SHARAD (Bramson et al., 2015) and recent ice-exposing impacts and scarps (Dundas et al., 2021) having indicated subsurface ice to be present. Additionally, in Appendix 1 we found that the glacial features largely lack crenulated terrain (i.e., brain terrain), which appears to be much more prominent in the surrounding terrain (Fig. A1.2). These differences in surface and near-surface properties suggests the buried ice sheet in the region varies laterally and vertically, and is therefore spatially inconsistent.

It is important to understand how the buried ice varies spatially and what implications this has for ice accessibility. As noted above, current orbital datasets cannot resolve the upper ~1‒10 m of the subsurface. However, surface morphology may be able to provide additional insight into the properties of the upper 10 m.

5.1. Brain Terrain Analogue

Brain terrain occurs in the mid-latitudes of Mars and is characterized by an anastomosing complex of ridges and troughs arranged in a “brain-like” pattern (Figs. 1.7, A1.3, and 5.1). There have been only a handful of studies on brain terrain (e.g., Bina and Osinski, 2021; Cheng et al., 2021; Levy et al., 2009; Noe Dobrea et al., 2007), and its relationship with ice is debated. Yet, brain terrain has been used as a morphologic indicator of buried ice (e.g., Fassett et al., 2013) and significant sublimation (e.g., Williams et al., 2017). In order to use brain terrain as a morphologic indicator of subsurface content, we must first understand how it forms and what relationship, if any, with ice it may have.
Levy et al. (2009) proposed that brain terrain forms in glacial ice deposits via thermal contraction cracking and subsequent topographic inversion from the sublimation of underlying ice (Fig. 5.2). However, Noe Dobrea et al. (2007) proposed brain terrain may form via frost heaving of clays (i.e., the expansion of soils via hydration) similar to the formation of stone circles on Earth. More recently, Cheng et al. (2021) proposed that brain terrain may form via aeolian processes and salt weathering/evaporite processes in the absence of ice.
Figure 5.2. Proposed brain terrain formation mechanism of Levy et al. (2009). (1) sublimation of underlying glacial ice forming a lag deposit, followed by (2) thermal contraction cracking exposing the underlying ice causing (3) the ice to sublimate and the crack to fill with sediment from the lag deposit and cement with pore-filling ice. The thin lag deposit provides less insulation to the underlying ice causing (4) the topography to become inverted via sublimation and creates raised ridges like closed cell brain terrain. The (5) flow of ice deforms the inverted cells creating a sinuous, elongate or arcuate nature. If the inverted ridges are ice cored, (6) the sublimation and subsequent collapse of these ice cores forms the flat floor of open cell brain terrain.

Brain terrain morphology is quite variable across Mars and, as a result, morphometrics also vary. Levy et al. (2009) separated brain terrain in Utopia Planitia into two categories
based on surface texture. This includes closed-cell brain terrain, which exhibits arcuate and mounded cells with flat or rounded upper surfaces, and open-cell brain terrain, which exhibits arcuate, elongate, or circular cells with a convex-up ridge sounding a flat-floored depression (Fig. 5.3). Ridges of closed cell brain terrain are reported to be 10–20 wide, 10–100 m long and 2–5 m high, and ridges of open cell brain terrain are commonly 4–6 m wide and 2 m high (Levy et al., 2009). Boulders were found to commonly reside on the top of brain terrain ridges, as well as in the troughs, and many ridges also exhibit cracks along the axial trace of the ridge (Fig. 5.3).

Cheng et al. (2021) provided morphometric measurements of the ridges of brain terrain from two HiRISE images (one in the northern hemisphere and one in the southern hemisphere). They found that brain terrain ridges in the northern hemisphere had a length between ~29–95 m, averaged at 52 m, and a width between ~8–12 m, averaged at 16 m. Brain terrain ridges in the southern hemisphere had a length between ~13–54 m, averaged at ~25 m, and a width between ~5–10 m, averaged at ~7 m. Height was not measured.

Brain terrain at our study site in Arcadia Planitia experiences a range of morphology (Fig. A1.3). However, open (Fig. 5.4) and closed cell brain terrain can be found across the study site. Brain terrain ridges in HiRISE image ESP_016465_2190 were measured to have a width ranging between 2–8 m.
Figure 5.3. Open cell brain terrain (OC-BT) and closed cell brain terrain (CC-BT). These occur on the same deposit on an undulating surface (rough topographic profile in white). Black specs can be found across the deposit which represent boulders. Yellow arrows indicate cracks along axial trace of ridges. From Levy et al. (2009).
We recently identified a landform, referred to Vermicular Ridge Features (VRFs), in the Canadian High Arctic on Devon Island (Chapter 3) and Axel Heiberg Island (Chapter 4), Nunavut, that has strikingly similar morphological characteristics to Martian brain terrain (Figure 5.5 and 5.6). VRFs exhibit a circular, elongate, sinuous and/or anastomosing ridge and trough morphology in planform (Figs. 3.3, 3.4, 4.3, 5.5, and 5.6) much like brain terrain (Figs. 5.5 and 5.6). Circular and semi-circular ridges with a central trough can be found in VRFs (Fig. 5.5a) and brain terrain (Fig. 5.5d). Circular ridges without a central trough can also be found on both VRFs (Fig. 3.3c) and brain terrain (Fig. 5.5d). Ridges exhibit a cracking along the axial trace and cobbles and boulders can be found resting atop ridges of VRFs (Fig. 5.6a) and brain terrain (Fig. 5.6b). Additionally, VRFs are only observed on flat-lying or gently sloping surfaces, and, to the authors knowledge, brain terrain has only been observed on flat-lying and gently sloping surface as well.

VRFs and brain terrain also have comparable morphometrics. VRF ridge width ranges between 1.5 and 9 m (more commonly around 2–4 m) and ridge height ranges between 0.2 and 2.5 m (more commonly around 0.5–1 m). This is comparable to morphometrics reported by Levy et al. (2009) and at Arcadia Planitia. VRFs can extend long lengths (Fig. A3.2) comparable to the lengths Cheng et al. (2021) observed. However, morphometrics in VRFs appear to vary across Axel Heiberg Island (Fig. 4.8).
Figure 5.5. VRF and brain terrain morphologic comparison of closed circular ridges. (a) Dundas Harbour VRFs exhibiting a sinuous, circular and anastomosing morphology. Yellow arrow points to crack along axial trace of ridge. Red box is panel b. (b) Circular (green arrow) to semi-circular (white arrow) ridges with central troughs. (c) Brain terrain from Levy et al. (2009) exhibiting a circular to semi-circular morphology. Red box is panel d. HiRISE image PSP_002175_2210. (d) Circular (green arrow) to semi-circular (white arrow) ridges with central troughs. Circular ridges without a central trough (orange arrow).
Both brain terrain and VRFs have also been observed to have a close spatial relationship with polygonally patterned ground (Fig. 5.7). However, it is unknown whether this is genetic. In Chapters 3 and 4 we observed thermal contraction cracks cross-cutting brain terrain features (Figs. 3.3, 3.4, 4.3, 4.4, and 4.8) indicating that this was a result of post-depositional periglacial modification. We suggested that ongoing thermal contraction cracking and subsequent wedge growth would raise the polygon shoulders which are composed on the same VRF deposit material (Fig. 5.7a). It is also possible frost-heave processes may contribute to further sorting along the polygon shoulders. In areas where thermal contraction cracking is minimal, VRFs appear closer together and the deposit may be more continuous as it is less disturbed (e.g., Fig. 5.7a). A similar relationship can be observed in brain terrain (e.g., Fig. 5.7b). However, Levy et al. (2009) suggests brain terrain and thermal contraction cracking are genetic. It is unclear how thermal contraction cracks this would affect the continued production of VRFs. However, we suggest thermal
contraction cracks are a post-depositional feature that reworks the original morphology of the VRF deposit.

Figure 5.7. VRF and brain terrain morphologic transition comparison and relationship with polygons. (a) VRFs at Axel Heiberg Island transitioning from clustered raised ridges (top right) to closed cell ridges with central troughs (middle) to polygons with raised shoulders (bottom left). VRF imagery using Esri (2018). World Imagery Source: Esri, Maxar, GeoEye, Earthstar Geographics, CNES/Airbus DS, USDA, USGS, AeroGRID, IGN, and the GIS User Community. (b) Brain terrain on Mars transitioning from raised circular closed cell ridges (bottom left) to open cell raised ridges encircling a central trough (middle) to high-centered polygons (top right). HiRISE image PSP_002175_2210.

We interpret VRFs to be remnants of ice-cored hummocky moraines (Chapter 3) or kame terraces (Chapter 4) that formed from the ablation and disintegration of buried glacial ice (Fig. 3.12). These are ice-marginal landforms which results in cobble-rich poorly sorted deposits overlying ice. The amount of ice preserved in the deposit varies and depends on the amount of ice initially buried, the thickness of the overlying sediment, and local environmental conditions. VRFs at Dundas Harbour on Devon Island (Chapter 3) showed no apparent buried massive ice remaining. However, it is unclear whether buried ice is currently present within the deposit. On Axel Heiberg Island (Chapter 4), VRFs were observed to be overlying massive ice at thaw slump exposures. Massive ice was estimated to exist ~2–6 m below the surface. The thickness of the deposit that makes up
the VRFs (regardless of the presence of ice) appeared to be ~10‒15 m thick. Periglacial activity has since altered the original deposit, resulting in cracking, solifluction, and slumping of material, which contemporaneously occurs with continued paraglacial modification (i.e., ice disintegration).

Based on the observations of VRF formation on Earth, we suggest that brain terrain on Mars forms from the disintegration of buried massive glacial ice, likely predominantly via sublimation. This would suggest buried ice is, or was, present at one point in time where brain terrain exists on Mars. However, massive ice cannot be inferred as presently extant from the identification of brain terrain alone. Additionally, we suggest thermal contraction cracking is not necessarily responsible for the formation of brain terrain, as suggested by Levy et al. (2009), but is a product of periglacial modification of the deposit that occurs concurrently with brain terrain formation via ice ablation.

However, further studies on VRFs are necessary to assess how appropriate of analogue they are for brain terrain on Mars. For example, an in-depth fabric and sedimentologic analyses would be useful for appropriately classifying VRFs as was done by Knudsen et al. (2006) on similar landforms. It may be also useful to investigate the mechanism and rate of buried ice degradation, and, thus, VRF formation, at VRF sites, much like McKenzie (1969) did at a collapsing kame terrace in Alaska. This could be done in part by collecting soil temperature data across various point along ridges and troughs over time to determine areas most vulnerable to thermal erosion. This, coupled with grain size analyses and moisture content, could be used to estimate the thermal conductivity of the deposit. These data could be used to model degradation under varying pressure and temperature conditions, including under current and possible past Martian conditions. Additionally, further mapping of VRFs across the Canadian High Arctic would be useful in determining the reach of this landform and assessing the various depositional environments and mechanisms in which it can form (i.e., kame terrace, ablation moraine, or other).
5.2. Implications for Arcadia Planitia

Brain terrain (i.e., crenulated terrain) is distinctly less prevalent on the glacial units identified in Chapter 2, particularly unit 1 (Figs. 2.2 and A1.2). However, brain terrain is suggested to represent a lag deposit overlying massive glacial ice (Levy et al., 2009). Brain terrain at our study site in Arcadia Planitia exhibits thermally bright properties (Appendix 1) that account for the moderate thermal inertia values observed in the surrounding terrain in Chapter 2 (Table 2.2). These relatively high thermal inertia values could represent a thin lag deposit overlying massive ice. However, a high stone concentration on the surface, covered by a thin layer of dust could produce the relatively high thermal inertia values as well. In this scenario, ice could still exist in the subsurface but may be too deep for GRS, THEMIS and TES to detect.

The presence of brain terrain at Arcadia Planitia likely indicates widespread buried massive ice exists but is undergoing or underwent near-surface ice ablation via sublimation. The depth to the ice is unclear but, based on VRF observations, would be greater than the brain terrain ridge height (> ~2 m based on Levy et al. (2009) observations). Therefore, the glacial units mapped (Fig. 2.2) in Chapter 2, particularly unit 1, likely represent areas where less near-surface ice ablation and disintegration has occurred and ultimately where ice is shallower from the surface. The glacial units from Chapter 2 are predominantly characterized by a polygonal surface morphology.

Therefore, more studies focusing on the relationship between polygon morphology and subsurface ice is needed to assess the depth and concentration of ice at the units mapped in Chapter 2.

Additionally, the presence of potential ice streams on Mars suggests more complex glacial dynamics existed than was previously thought likely. Ice streams on Earth move rapidly as a consequence of their beds being lubricated by liquid water which has major implications for past climate and former rudimentary life on the planet. Our understanding of subglacial hydro-mechanical processes on Mars is very limited and debated. This interpretation, however, is challenging, nascent and requires corroboration and refinement. Modeling of ice flow dynamics under different climatic and subsurface
conditions would be useful for testing this interpretation. Nonetheless, flowing streams of ice in a flat-lying region on Mars has not been previously observed. The dynamics involved in the formation, flow and preservation of this ice provides scientific value to a potential future mission to Arcadia Planitia as it has implications for past climate and, if water were to be present, habitability.

Therefore, we conclude that the glacial units mapped in Chapter 2, particularly unit 1 where potential ice streams exist, represent favourable regions for ISRU and scientific value for future human exploration.

5.3. References


Appendix 1: Mapping Near-Surface Ice Surface Morphologies in Arcadia Planitia, Mars

A1.1. Introduction

Near surface ice on Mars becomes stable at mid-latitudes during high obliquity periods (>30°) (Head et al., 2003). Widespread desiccation of mid-latitude subsurface ice via sublimation occurs during low obliquity periods and ice becomes stable at the poles (Head et al., 2003). Mars’ obliquity has ranged between 15° and 35° over the past 3 Myr (Laskar et al., 2004) which may have led to the deposition of mid-latitude ice. However, over the past 300 kyr the obliquity has remained moderate ranging between 22° and 26° leading to the desiccation and degradation of equatorial ice (Head et al., 2003). Evidence of an ice-rich latitude-dependent mantle (LDM), an ice-rich deposit blanketing much of the mid to polar latitudes, has been suggested poleward of 30°; however, ice is less stable at mid-latitudes under current obliquity and climate conditions.

Arcadia Planitia is a smooth plain with small knobs located at ~200°E, ~40°N (~160°W, ~40°N) and is a low-lying region on Mars where significant near-surface water ice is suggested to be present. Gamma ray spectroscopy reveals a H subsurface ice layer makes up 35 ± 15% by weight, most of which is found poleward of 40°N and -40oS (Boynton et al., 2002). Evidence of debris-covered lobate flows have been identified in western Arcadia (Mangold et al., 2003; Plaut et al., 2009). Orbital imaging has revealed recent small impacts that have exposed subsurface ice at mid- and high-latitudes (Byrne et al., 2009; Dundas et al., 2014). Thermokarstic expansion of secondary craters indicates ground ice sublimation (Viola et al., 2015). Concentric terracing within the walls of simple craters indicates layers of ice within the target material (Bramson et al., 2015). Shallow Radar (SHARAD) detects a dielectric constant similar to that of ice (Holt et al., 2008; Plaut et al., 2009; Bramson et al., 2015). Both pore-filling and excess ice have been found within the upper cms of soil at the Phoenix landing site at 68°N, 234°E (Mellon et al., 2009; Smith et al., 2009; Arvidson et al., 2009). Polygonal surface patterns, indicative of cryoturbation and ground ice sublimation, were identified in Arcadia, at Phoenix, and
at higher latitudes (Mellon et al., 2009; Levy et al., 2009a; Williams et al., 2017). HiRISE and CTX images reveal morphological indications of near surface ice (Williams et al., 2017), including (1) polygonal terrain, which consists of small mounds encircled by polygonal troughs, interpreted to be the result of thermal contraction (Arvidson et al., 2009; Mellon et al., 2009; Levy et al., 2009a), (2) crenulated terrain (i.e., brain terrain), which consists of sinuous ridges and troughs, interpreted to be the result of thermal contraction and differential sublimation (Noe Dobrea et al., 2007; Levy et al., 2009b), and (3) pitted terrain, which consists of shallow depressions, interpreted to be the result of localized sublimation along thermal contraction fractures (Mangold, 2011; Williams et al., 2017). A latitude dependent transition in morphology indicates the desiccation of the LDM preferentially at lower latitudes (Fig. A1.1; Williams et al., 2017).

Figure A1.1. Latitudinal distribution of surface morphology frequency in Arcadia Planitia. The greatest ice loss occurs in the middle latitudes of the study site where the greatest crenulated terrain exists. From Williams et al. (2017).

The formation of excess ice in the LDM and understanding its persistence at mid-latitudes is still unknown. In order to facilitate LDM studies and the identification of
spatial relationships in Arcadia, the goal of this study is to expand on the findings of Williams et al. (2017) and identify contiguous spatial relationships of ice-related morphologic indicators to infer abundance and LDM dynamics. This study aims to answer three step-wise questions, including (1) What are the distributions of the ice-related morphologies, (2) Are ice-related morphologies latitude-dependent, and (3) Is there a poleward regression of LDM ice? These results will aid in understanding why mid-latitude ice still exists. We aim to create an ice-related historical series of events in the Arcadia Planitia region to assist in climate modeling of mid-latitude ice and for the identification of a future landing site for human missions and ISRU.

A1.2. Methods

A1.2.1. Georeferencing

A Thermal Emission Imaging System (THEMIS) daytime infrared (IR) mosaic (100 m/pixel) was the highest resolution basemap available for Arcadia Planitia. Mars Reconnaissance Orbiter (MRO) Context Camera (CTX) images (6 m/pixel) were initially georeferenced and superimposed onto the THEMIS basemap by hand using ArcMap. Obvious features, such as mesas or craters, are used to create tie-points, which superimpose one image over another accurately. Craters are the most useful feature because they have a constant diameter making it easier to accurately select tie-points. Up to 80 tie-points can be necessary to create an accurate georeferenced image. However, georeferencing by hand was no longer necessary when an automated georeferencing system had been developed.

A1.2.2. Coding

A total of 38 CTX images were used to create a basemap ranging between around 38°N to 44°N latitude and around 155°W to 167°W. Mars Nest, a software implemented within Automated Fusion of Image Data System (AFIDS), is an automated georeferencing system on Linux that has recently been developed at the Jet Propulsion Lab (JPL) and has been used in place of ArcMap for georeferencing. Mars Nest follows a particular set of parameters set by the user to identify patterns to align images and creates over 1000 tie-
points. This project has been used to configure and improve the software for future use.

CTX images were georeferenced onto the THEMIS basemap using Mars Nest with a fast fourier transform (fft) window size of 128, fftgrid of 64 x 64, which sets the number of pixels in the moving window, and a magnification of 1-1, which looks pixel-by-pixel. A CTX mosaic was assembled in ArcMap to create a basemap of CTX images as they are available and cover all of Arcadia. However, the resolution is not high enough for describing the geomorphology.

MRO’s High Resolution Imaging Science Experiment (HiRISE) images (0.25 m/pixel) have a much higher resolution that can be used to identify morphologic indicators of near surface ice, such as polygonal, crenulated, and pitted terrains. A total of 72 red-filter HiRISE images (0.25m/pixel) of Arcadia were georeferenced to the CTX basemap using Mars Nest with an fftsize of 256, fftgrid of (64,64), a magnification of 2-1, which skips every other pixel to cover a wider area, and a border of 1000, which adds margin to the image to prevent clipping. All images overly the CTX basemap and were assembled into a mosaic in ArcMap. Mars Nest was able to produce a semi-accurate (<200 m) CTX basemap, as well as HiRISE footprints. HiRISE images were used to interpret and interpolate features across the more widespread lower resolution CTX images

### A1.2.3. Mapping

HiRISE images were used to identify morphologic indicators of near surface ice. Surface morphologies are representative of the geologic process occurring. Three main categories of near surface ice morphologies have been identified, including polygonal, crenulated, and pitted. These images also assisted in the interpretation of morphologies on the lower-resolution CTX basemap. Each morphology is mapped in ArcMap using a shapefile and drawing multiple polygons manually.

### A1.3. Observations

A map of ice-related surface morphologies was created for Arcadia Planitia (Fig. A1.2). Three major units (polygonal, crenulated, and pitted), and three transitional units have been mapped. Variations, or subtypes, occur within the major units (Table A1.1; Fig.
Figure A1.2. Map of ice-related surface morphologies in Arcadia Planitia overlying THEMIS daytime IR mosaic. Three primary surface morphologies (polygonal, crenulated, and pitted) and three transitional surface morphologies (lineated polygonal, weak crenulated, and weak pitted) are mapped. Additional variations exist within the three primary morphologies (Fig. A1.3). Polygonal terrain is the most expansive unit.
Table A1.1. List of three primary surface morphologies in Arcadia Planitia and their variations (i.e., subtypes).

<table>
<thead>
<tr>
<th>Type</th>
<th>Subtype</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polygonal</td>
<td>typical</td>
<td>10-15 m diameter; distinguishable troughs; found everywhere</td>
</tr>
<tr>
<td></td>
<td>lineated</td>
<td>typical polygons with linear features preferentially oriented (typically to the SE); appears bright in CTX</td>
</tr>
<tr>
<td></td>
<td>basketball</td>
<td>~10 m diameter; high centered polygons; typically seen at higher latitudes</td>
</tr>
<tr>
<td></td>
<td>distinct</td>
<td>10-15 m diameter; distinct polygon troughs compared to typical polygons</td>
</tr>
<tr>
<td></td>
<td>weak</td>
<td>10-15 m diameter; subdued or indistinct troughs; typically seen at lower latitudes</td>
</tr>
<tr>
<td>Crenulated</td>
<td>typical</td>
<td>interconnected sinuous ridges and troughs; appear bright in visible and IR; tend to occur at middle latitudes</td>
</tr>
<tr>
<td></td>
<td>linear</td>
<td>sinuous ridges and troughs linked together in a line all having a preferred orientation; tend to occur at the middle latitudes</td>
</tr>
<tr>
<td></td>
<td>bumpy</td>
<td>clusters of bumps and ridges with pointed edges; typically seen at higher latitudes</td>
</tr>
<tr>
<td></td>
<td>weak</td>
<td>sinuous ridges and troughs are isolated or clustered; individual circular or crescent-shaped ridges; typically found as a transition from crenulated or polygonal terrain</td>
</tr>
<tr>
<td>Pitted</td>
<td>typical</td>
<td>polygonal terrain with abundant pitted troughs; troughs preferentially aligned (typically to the SW); found in the NE</td>
</tr>
<tr>
<td></td>
<td>weak</td>
<td>fewer pitted troughs present; preferentially aligned (typically to the SW)</td>
</tr>
</tbody>
</table>
Figure A1.3. Variations (i.e., subtypes) among the three primary surface morphologies (polygonal, crenulated, and pitted). Descriptions provided in Table A1.1. (a) Typical
Polygonal terrain. Image located at 39.5°N, 162°W. HiRISE image ESP_048086_2200. 
(b) Lineated polygonal terrain. Image located at 40.8°N, 161°W. HiRISE image 
ESP_025682_2210. (c) Basketball polygonal terrain. Image located at 42.5°N, 162.4°W. 
HiRISE image ESP_033818_2230. (d) Distinct polygonal terrain. Image located at 42°N, 
165.7°W. HiRISE image ESP_015898_2225. (e) weak polygonal terrain. Image located 
at 38.7°N, 156.9°W. HiRISE image ESP_050156-2190. (f) Typical crenulated terrain. 
Image located at 39.5°N, 157°W. HiRISE image ESP_025339_2195. (g) Linear 
crenulated terrain. Images located at 39.7°N 157.1°W. HiRISE image 
ESP_049589_2200. (h) Bumpy crenulated terrain. Image located at 42.4°N 162.4°W. 
HiRISE image ESP_033818_2230. (i) Weak crenulated terrain. Image located at 39.2N 
165°W. HiRISE image ESP_017243_2195. (j) Typical pitted terrain. Image located at 
42.8°N 158.3°W. HiRISE image ESP_034253_2230. (k) Weak pitted terrain. Image 
located at 41.9°N 159.2°W. HiRISE image ESP_034820_2225.

Polygonal terrain contains five sub-types (Table A1.1), including typical (Fig. A1.3a), 
lineated (Fig. A1.3b), basketball (Fig. A1.3c), distinct (Fig. A1.3d), and weak (Fig. 
A1.3e). General latitudinal observations have been made indicating weak polygons occur 
in the lower latitudes and basketball polygons or polygons with more distinct troughs 
occur in the higher latitudes of the mapping area. There are two major sinuous units with 
a wrinkled, knobby, polygonal terrain trending SE until they equivocally dissipate south 
of 38°N into weaker polygonal terrain. Polygonal terrain can have knobs (Fig. A1.3e), 
which becomes more prevalent in lower latitudes. The two sinuous units appear to act as 
a boundary for the crenulated and lineated terrain indicating a difference from its 
surroundings.

Crenulated terrain appears bright in visible and thermal IR. Crenulated terrain contains 4 
sub-types (Table A1.1), including typical (Fig. A1.3f), linear (Fig. A1.3g), bumpy (Fig. 
A1.3h), and weak (Fig. A1.3i). All crenulated brain terrain tends to occur in the middle 
latitudes of the mapping area and does not extend as far north or as far south as polygonal 
terrain in the region. However, a distinct morphological change in crenulated brain terrain
occurs around 41°N into bumpy crenulations only. Crenulated terrain occurs within a latitudinal band between ~ 38°N and 42°N. Secondary craters found in polygonal terrain around ~ 41.5°N have crenulations on the south-facing slopes of the crater walls (Fig. A1.4). Craters north of this latitude observed in available HiRISE images do not show any evidence of crenulated terrain. Alternating light and dark bands between 40.5°N and 41.5°N that are predominantly crenulated near the southern limit and become decreasingly crenulated northward (Fig. A1.4).

Figure A1.4. Secondary craters at 41.5°N, 159°W. HiRISE images overlying THEMIS daytime IR mosaic. Red boxes indicate panels b and c. (b) Secondary crater with polygonal terrain. Some knobs can be seen on the south-facing slope. HiRISE image ESP_034820_2225. (c) Secondary crater with polygonal terrain. Crenulated terrain (arrow) on south-facing slope. HiRISE image ESP_035598_2220.
Pitted polygonal terrain contains 2 sub-types (Table A1.1), including typical (Fig. A1.3j) and weak (Fig. A1.3k). Pits occur only within the troughs of polygons and are always oriented unidirectionally, always to the SW. Pitted polygonal terrain only occurs in the northeastern section of the study area. Typical pitted terrain tends to be dark in thermal IR. However, bright streaks in CTX found on the eastern edge of the map have crenulations at lower latitudes and pits at higher latitudes (Fig. A1.5).

Figure A1.5. Bright streaks in CTX exhibiting a change in morphology northward. HiRISE images overlying CTX mosaic. Red boxes indicate locations of panels b and c. (b) Green box showing location of panel d on CTX. (c) Green box showing location of panel e on CTX. (d) Weak crenulations on bright streak. HiRISE image PSP_006826_2215. (e) Pitted terrain on bright streak. HiRISE image ESP_034965_2220.

A1.4. Discussion

The lowest latitudes where widespread LDM ice exists is in Arcadia Planitia, located around 38°N-40°N (e.g. Pathare et al., 2018). Surface morphologies indicative of ice processes were mapped in the lower mid-latitudes of Arcadia Planitia. A diverse set of morphologies exist in the mapping region that are grouped into three main categories, including polygonal terrain, crenulated terrain and pitted terrain.
Levy et al. (2009a) found a latitudinal dependence in polygon morphology on Mars. They suggest older polygonal surfaces exist at lower latitudes and younger polygonal surfaces exist at higher latitudes. They suggest that high center polygons indicate initial stages of polygon development, whereas low center polygons indicate advanced stages of polygon development with localized sublimation at polygon centers. High center polygons may represent sublimation polygons indicating underlying excess, or even massive, ice.

Evidence of desiccation at the edge of the LDM suggested by Williams et al. (2017) is observed by the disposition of the surface morphology. Lower latitudes (< 39°N) within the study area exhibit predominantly weak polygonal terrain, and some typical and weak crenulated terrain indicating significant degradation of the LDM via sublimation. Middle latitudes (39°–41°N) are characterized predominantly by typical and lineated polygonal terrain and typical, linear and weak crenulated terrain indicating increased sublimation of a degrading LDM. Higher latitudes (> 41°N) are dominated by basketball and distinct polygonal terrain, bumpy crenulated terrain, and pitted terrain indicating localized sublimation. Additionally, south-facing slopes of secondary crater walls show crenulations indicating increased insolation leads to increased sublimation. This transition in the spatial distribution of surface morphologies indicates a latitude-dependent transgression in ice stability. We interpret this to represent a poleward regression of the LDM.

These findings provide implications for both current and past climate and LDM studies, by (1) indicating that water-ice is unstable at mid-latitudes and is likely being relocated to and deposited at higher latitudes, and (2) that conditions in the past were more favorable to the deposition of water-ice at mid-latitudes and sublimation at the poles.

The suggestion of widespread ice present at lower latitudes holds significant implications for potential future in situ resource utilization. If humans colonize Mars, it is necessary they land nearest the equator and have access to a substantial amount of water-ice to use to consumption, fuel, and a sustainable agriculture. Additionally, detailed maps of small-scale surface morphologies may provide assistance in identifying a safe landing site and safe traversability.
A1.5. References


Mangold, N., 2003, Geomorphic analysis of lobate debris aprons on Mars at Mars Orbiter Camera scale: Evidence for ice sublimation initiated by fractures: Journal of


Figure A2.1. Sinuous features S1, S2, and S3 in Arcadia Planitia are highlighted in purple on THEMIS Daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Major tributaries and southernmost region connecting sinuous features has been removed for individual sinuous feature data collection presented in Table 2.1.
Figure A2.2. Map of glacial-related features in the Eastern Erebus Montes region in Arcadia Planitia on THEMIS Daytime IR mosaic (Edwards et al., 2011; Hill and Christensen, 2017). Appendix 2 figure locations are boxed in white and labeled for context.

<table>
<thead>
<tr>
<th>Legend</th>
<th>Unit Description</th>
<th>Unit Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>Sinuous, thermally bright features with a polygonal, ridged and knobby surface texture and arculate features</td>
<td>Buried massive ice deposit with a glacial-related origin akin to lineated valley fill</td>
</tr>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>Lobate features with subdued polygons and near-linear down-slope ridges</td>
<td>Lobate debris aprons</td>
</tr>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>Rounded cratered hills with boulders</td>
<td>Massifs/exposed bedrock</td>
</tr>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>Lobate, gently sloped aprons with near-concentric deformed ridges</td>
<td>Ice-cored lobate debris apron moraines</td>
</tr>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>Thermally dark mantling unit with extensive brain terrain and linear ridges</td>
<td>Ice-rich mantling unit</td>
</tr>
<tr>
<td><img src="image" alt="Legend Image" /></td>
<td>A smooth textured and polygonised terrain with pitted troughs and abundant expanded secondary craters</td>
<td>Transitional ice-rich mantling unit connecting all ice-rich units to one another</td>
</tr>
</tbody>
</table>

End of individual sinuous features for data collection
Figure A2.3. Morphologies observed on unit 2. Unit boundaries indicated in bottom left image in each panel for context. (a) Parallel linear forms at the base of unit 2 (red arrow) and expanded craters on unit 2 (white arrows). HiRISE PSP_009740_2200. (b) Expanded crater (white arrow) and crack running parallel to unit 2 and 3 boundary (red arrow). HiRISE ESP_026447_2200. (c) Cracks running parallel to unit 2 and unit 3 boundary (red arrows). HiRISE ESP_027436_2205. (d) Boundary between unit 3 and unit 6. No clear surface morphologic change is observed but a change in elevation indicates the boundary (white arrow). HiRISE ESP_027436_2205.
Figure A2.4. Morphologies observed on unit 3. Unit boundaries indicated in bottom left image in each panel for context. (a) Cratered surface (white arrow) of a rounded/convex-up unit 3 outcrop. Clear boundary between units 2 and 3 evident by small moat and distinct change in relative brightness (red arrow). HiRISE PSP_009740_2200. (b) Blocky surface of unit 3 with large boulders present (red arrow). HiRISE PSP_009740_2200. (c) Heavily cratered surface on a flat unit 3 outcrop. Boundary between units 2 and 3 (red arrow) evident by a change in crater density, difference in crater expansion, and scarp face. HiRISE ESP_030047_2200. (d) Parallel linear grooves and ridges (red arrow) aligned NE-SW on a flat unit 3 outcrop. HiRISE ESP_016465_2190.
Figure A2.5. Morphologies observed on unit 4. Unit boundaries indicated in bottom left image in each panel for context. (a) Deformed sinuous and near-concentric ridges (white arrow) and pits (red arrows). CTX B01_010162_2203_XI_40N166W. (b) Series of sinuous and anastomosing ridges and troughs consistent with crenulated, or brain, terrain. Boundary between units 2 and 4 is observed by a sudden change in morphology from smooth (unit 2) to textured (unit 4) (red arrow). HiRISE ESP_017243_2195. (c) Patchy variation of features consistent with crenulated, or brain, terrain on unit 4. HiRISE PSP_009740_2200. (d) Boundary between units 4 and 6 indicated by distinct changes in morphology, such as the presence/absence of crenulated/brain terrain (blue arrow), parallel linear features along a gentle slope (red arrow), and a change in relative brightness (green arrow). HiRISE ESP_016465_2190.
Figure A2.6. Morphologies observed on unit 5. Unit boundaries indicated in bottom left image in bottom two panels for context. (a) Subdued parallel linear forms on unit 5. Direction indicated by red arrow. HiRISE ESP_017243_2195. (b) Prominent parallel linear forms on unit 5. Direction indicated by red arrow. HiRISE PSP_009740_2200. (c) Boundary between units 4 and 5 identified by extent of deformed ridges and lobate nature of unit 5 (red arrow). Change in relative brightness is apparent at unit 5 and surround terrain boundary (white arrow). This boundary can also be observed in THEMIS daytime IR (See Figure 2.1b). CTX D22_035796_2198_XN_39N164W. (d) Boundary between unit 5 and surrounding terrain can be identified by a lobate feature and change in relative brightness (red arrow). This boundary can also be observed in THEMIS daytime IR (See Figure 2.1b). CTX K20_061075_2195_XN_39N164W.
Figure A2.7. Morphologies observed on unit 6. Unit boundaries indicated in bottom left image in each panel for context. (a) Boundary between units 1 and 6 can be gradual and often unclear but can be identified by a change in surface morphologies. The ridged and bumpy surface morphologies on unit 1 (red arrow) are not present in unit 6 and a slight change in relative brightness can be observed. Expanded secondary craters are abundant in both units 1 and 6 (white arrows). CTX K19_060719_2201_XN_40N164W. (b) Boundary between unit 6 and surrounding terrain is identified by a change in surface morphology (red arrow). Unit 6 is smooth, flat, dark and has many expanded craters (white arrow). HiRISE ESP_016465_2190. (c) Boundary between units 4 and 6 evident by a change in relative brightness, morphology, and extent of deformed ridges (red arrow) Expanded craters can be commonly found in unit 6 (white arrow). HiRISE ESP_017243_2195 (d) Boundary between unit 6 and surrounding terrain is gradual. This boundary can be identified by a change in relative brightness and morphology (red arrow). This boundary can be patchy and exposed within unit 6 (blue arrow). Small patches out surrounding terrain
are included in unit 6. Many expanded craters (white arrow) can be found on unit 6 and disappear in the adjacent terrain (red arrow). HiRISE ESP_030047_2200.
Appendix 3: Supplementary Materials for Chapter 3

a

b
Figure A3.1. Examples of cracking along ridges of VRFs. (a) Example of thermokarst processes creating topographic depressions and ridges in the diamicton layer in Site 2. (b) Example of cracks along the axial trace of a ridge where no evidence of thermokarst processes are evident in Site 1.
Figure A3.2. Examples of elongated VRFs traced in white. (a) Longest example of a closed cell vermicular ridge feature at Site 1 spanning 72 m across perpendicular to slope. (b) Longest stretching vermicular ridge feature at Dundas Harbour spanning 120 m directly across (curve length not included) and located in Site 3. This ridge does not create a circular or closed cell and marks the end of the deposit at Site 3 where it overlies raised beach deposits. Ridge runs perpendicular to slope.
Appendix 4: Footnotes

1 A version of this chapter/appendix has been published in Icarus (Hibbard, S.M., Williams, N.R., Golombek, M.P., Osinski, G.R. and Godin, E., 2021. Evidence for widespread glaciation in Arcadia Planitia, Mars. Icarus, 359, 1, 114298. https://doi.org/10.1016/j.icarus.2020.114298). Authors can use their articles, in full or in part, for a wide range of scholarly, non-commercial purposes one of which is inclusion in a thesis or dissertation. Elsevier’s policy is to allow authors to use their work in their thesis or dissertation that can be made publicly available on the author’s university web site/repository for non-commercial use (https://www.elsevier.com/about/our-business/policies/copyright/personal-use).

2 A version of this chapter/appendix has been submitted for publication in Geomorphology (Hibbard, S.M., Osinski, G.R. and Godin, E., In Review. Vermicular Ridge Features on Dundas Harbour, Devon Island, Nunavut. Geomorphology). Authors can use their articles, in full or in part, for a wide range of scholarly, non-commercial purposes one of which is inclusion in a thesis or dissertation. Elsevier’s policy is to allow authors to use their work in their thesis or dissertation that can be made publicly available on the author’s university web site/repository for non-commercial use (https://www.elsevier.com/about/our-business/policies/copyright/personal-use).

3 Chapter 4 is undergoing preparation for submission for publication.

4 Appendix 1 is currently formatted as a short report that will be written up into a full paper in the future.
Curriculum Vitae

Name: Shannon Hibbard

Post-secondary Education and Degrees:

The Ohio State University
Columbus, Ohio, USA
2010-2014 B.S.

Temple University
Philadelphia, Pennsylvania, USA
2015-2017 M.S.

The University of Western Ontario
London, Ontario, Canada
2017-2021 Ph.D.

Honours and Awards:

Europlanet 2024 RI Programme & Transnational Access Grant
Fieldwork Grant
2021-2022

Mitacs Research Training Award (RTA)
Doctoral Fellowship
2020

Western Summer 2020 Student Internship Award
Scholarship
2020

International Association of Cryospheric Sciences Travel Grant
2020

Centre for Planetary Science and Exploration (CPSX) Travel Grant
2018

Related Work Experience:

Teaching Assistant
The University of Western Ontario
2017-2021
Dissemination Committee
Canadian Permafrost Association
2021

Glacial and Quaternary Geology Instructor
2018

Exploring Earth Blog Writer
Canadian Space Agency, The University of Western Ontario
2018

Centre for Planetary Science and Exploration Outreach Officer
The University of Western Ontario
2018

Mission Experience:

HiRISE Science and Operations Planning Volunteer
Dr. Livio Tornabene, The University of Western Ontario
2017-2021

Observer
Europa Clipper Project Science Group #9
2020

Science-Planning Investigator
Lunar Analogue CanMoon mission
The University of Western Ontario
2019

HiRISE Science and Operations Planning Cycle 316 Participant
Dr. Livio Tornabene, The University of Western Ontario
2018

Student Mentorship Experience:

Mentor of Student Intern
The University of Western Ontario
Western BSc Wendy Boucher
2021

Mentor of Student Intern
The University of Western Ontario
Western BSc student Robert Silber
2021
Mentor of Student Intern
The University of Western Ontario
Western BSc student Gabriela Robles Ospina
2021

Mentor of Student Intern
The University of Western Ontario
Western BSc student Alyssa Coelho
2020

Mentor of Student Intern
The University of Western Ontario
Western BSc student Rhiannon Punch
2019

Mentor of Student Intern
The University of Western Ontario
Western BSc student Anthony Dicecca
2019

Mentor of Online Research Co-op Program Participant
High School student Aodhán Corrigan
2019

Field Experience:

Arctic Conducted field work for my Ph.D. thesis regarding periglacial Earth-Mars analogues on Devon Island and Axel Heiberg Island, Nunavut, Canada. 2018, 2019

Sudbury Participated in field school with University of Western Ontario and LPI regarding impact processes, complex crater formation and ejecta distribution. 2017

Arizona Participated in field school with LPI regarding impact processes, simple crater formation and ejecta distribution at Barringer crater. 2016

South Africa Conducted field work for my Masters thesis regarding Precambrian impact ejecta produced by impact-derived vapor plumes by collecting samples containing spherules in Barberton, South Africa. 2016

Antarctica Assisted in research regarding ancient buried ice and paleo-landscape change by performing GPR surveying across rugged terrain, and collecting soil and ice samples in Garwood Valley, Taylor Valley, and Victoria Valley, Antarctica 2014
Austria Assisted in research regarding proglacial sediment budget and landscape change in response to climate change by stream gaging and using LiDAR in Feichten im Kaunertal, Öztal Alps, Austria. 2013, 2014

Norway Assisted in research regarding rockfall age-dating using a dating tool called the Schmidt Hammer in Jotunheimen National Park, Norway. 2014

New Zealand Assisted in research regarding end moraine age-dating using a dating tool called the Schmidt Hammer in Mount Cook National Park, New Zealand. 2014

San Salvador Participated in a course regarding carbonate depositional environments on San Salvador Island, Bahamas. 2013

Utah Participated in field camp with The Ohio State University in Ephraim, Utah. 2012

Publications:


relevant conference abstracts:


**Highlights in the News:**

CBC News  
Newly discovered glaciers on Mars may help humans settle on the Red Planet one day

Western University  
Newly discovered ‘glaciers’ could aid human survival on Mars

Tom McConnell Show  
Shannon Hibbard, Glaciers on Mars

CBC Canada Tonight  
Newly discovered underground glaciers on Mars... right where we want them

LiveScience  
Possible new type of glacier just discovered on Mars

New York Daily News  
Study: Newly revealed Mars glacier could support human exploration

GlacierHub  
Can a New Type of Glacier on Mars Aid Future Astronauts?

Western University  
Western planetary scientists assist in capturing first full-colour image of NASA InSight using HiRISE space camera