

**Cold-Climate (Periglacial) Landforms on the Earth and Mars:
Geomorphic Evidence for Ice-Related Flow and Conditions for the
Generation of Meltwater.**

by

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A Dissertation Submitted in Partial Fulfillment of the

Requirements for the Degree of Doctor of Philosophy

in the Department of Geological Sciences at Brown University

Providence, Rhode Island

May 2009

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This dissertation by Gareth A. Morgan is accepted in its present form
by the Department of Geological Sciences as satisfying the
dissertation requirement for the degree of Doctor of Philosophy.

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Involved in a range of original research projects concerned with investigating climatic changes on Mars and the Earth through the use of remotely sensed data and fieldwork in terrestrial analog environments. Projects have included: 1) Geomorphological assessment of potential Amazonia/Hesperian glacial deposits and small-scale fluvial features on the surface of Mars. 2) In-depth analysis of fluvial erosion within the McMurdo Dry Valleys and its relationship to variations in solar insolation.

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Morgan, G.A., Head, J.W., 2009. Sinton crater, Mars: Evidence for impact into a plateau icefield and melting to produce valley networks at the Hesperian-Amazonian boundary. *Icarus*, 10.1016/j.icarus.2009.02.025.

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Head, J.W., Murchie, S.L., Prockter, L.M., Solomon, S.C., Chapman, CR., Strom, R.G., Watters, T.R., Blewett, D.T., Gillis-Davis, J.J., Fassett, C.I., Dickson, J.L., **Morgan, G.A.**, Kerber, L., 2009. Volcanism on Mercury: Evidence from the first MESSENGER flyby for extrusive and explosive activity and the volcanic origin of plains, *Earth and Planetary Science Letters*, In press.

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Involvement in Planetary Missions.

During my time at Brown University, I have participated in both the recent **Mercury 1 flyby** (Jan ’08, Applied Physics Laboratory, MD) by the **Mercury Surface, Space Environment, Geochemistry, and Ranging (MESSENGER)** spacecraft and the **High Resolution Stereo Camera (HRSC)** mission onboard **ESA Mars Express**. Through my involvement I have provided assistance to the science teams in the analysis and interpretation of the data sets.

Field Experience

Oct. 2006 – Jan. 2007. Oct. 2008 – Dec. 2008. McMurdo Dry Valleys, Antarctica. NSF expeditions. Brown University and Boston University.

Spent two field seasons as part of a small science team camped in the Antarctic Dry Valleys. Projects included: drilling and ice core retrieval from debris covered glaciers; establishing and operating seismic lines (in order to investigate subsurface properties); using ground probing radar; installing climate/environmental sensors; environmental sampling and monitoring, including ‘ground truth’ samples to calibrate satellite data; and undertaking geological reconnaissance and mapping expeditions.

Oct. 2003. Solent Estuary, UK. National Oceanography Centre.

Investigated the hydrographical properties of the estuarine water column through in situ sampling and the deployment of boat towed Conductivity, Temperature, Depth (CTD) devices.

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Investigated glacial debris transport and depositional landforms. Conducted detailed field research of glacial environments and operated ice-penetrating radar in order to map ice thickness.

Teaching Experience

2005 - Teaching Assistant. Brown University, Geological Sciences 05: Mars, the Moon and Earth. Provided assistance during lectures and taught labs involving the geologic/geomorphic mapping of planetary surfaces, the mineralogy of Apollo lunar samples and the exploration of the surface of Mars through the use of an immersive virtual reality facility.

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Taught classes on a range of subjects to 5th grade classes (10-11 years old).

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I have experience and proficiency in a range of software packages for the analysis and manipulation of quantitative data, including the use of the **ArcGIS** Geographical Information Systems software suite and **Matlab**.

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Preface

Under the current paradigm of Mars history, the planet is perceived to have undergone a global climatic change around 3.5 billion years ago, from relatively ‘warm and wet’ conditions (during the Noachian), conducive to the formation of dense valley networks and lakes, to ‘cold and dry’ conditions (during the Hesperian and Amazonian) characterized by temperatures and pressures largely below the triple point of water (e.g. Carr, 1996). Recent analysis of high-resolution image data have revealed an array of young landforms associated with both ice and liquid water at equatorial and mid-latitudes, where it is currently unstable (e.g. Malin and Edgett, 2000; Mustard et al., 2001; Head et al., 2005), suggesting significant climatic change has also occurred during the post-Noachian. Understanding the nature of the change is thus an important scientific goal and is essential to make an assessment of the potential for life to have existed over recent Mars history. The thesis presented here focuses on the investigation of the duration and spatial extent of post-Noachian climate transitions, through the morphological study of landscapes and forms indicative of the action of ice and liquid water. In taking this approach the thesis has drawn from, and contributed too, other Amazonian climatic investigations that have been presented in the literature (for example: Mustard et al., 2001; Milliken et al., 2003; Head et al., 2006a,b)

The thesis explores the post-Noachian climate in chronological order, and the first chapter opens with an investigation into Hesperian/Early Amazonian valley networks. The valleys surround the outer rim of the 60-km-diameter Sinton impact crater on the edge of the Northern Dichotomy Boundary within Deuteronilus Mensae. The discovery of ancient valley networks on the surface of Mars by the Mariner 9 mission (Masursky, 1973), provided the first major evidence of warmer and wetter conditions early in the planet’s history, as the presence of valleys argues for the occurrence of pluvial activity and prolonged surface runoff. This requires a thicker atmosphere with higher amounts of water vapor than is currently present on Mars (surface

pressure: one hundredth of Earth, water vapor: 1–100 μm : Farmer et al., 1977). However, valley networks can form in the absence of rainfall, through a range of processes including: spring sapping (Laity and Malin, 1985) and hydrothermal activity (e.g. Gulick and Baker, 1990; Dohm and Tanaka, 1999). Through a detailed morphological investigation we find that the formation of the Sinton valley networks is most consistent with the release of meltwater generated by the interaction between the impact that formed Sinton crater and preexisting surficial ice deposits. This model negates the need for pluvial activity, but does require the widespread atmospheric deposition of snow and ice. The chapter concludes that the Hesperian was punctuated by glacial activity, a scenario that is new and is consistent with evidence suggesting Hesperian/Early Amazonian valley networks on volcanic edifices were formed by the endogenic heating of surficial ice deposits (Fassett and Head, 2006; 2007).

Chapter Two continues the investigation of potential glacial activity along the Dichotomy Boundary by investigating Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA) within the Sinton crater region. These deposits were identified in Viking images and were initially interpreted to be mass-wasting products (talus piles) lubricated by the accumulation of ice in pore spaces permitting them to flow (Squires, 1978). However, a reevaluation of their morphology by Lucchitta (1984) argued that the deposits exhibited down-valley flow and were reminiscent of terrestrial glacial systems. Recent work by Head et al (2006a,b) suggested that LVF and LDA were the remnants of debris covered glaciers that contain significant volumes of glacial ice preserved from sublimation by the thin, dry atmosphere through the development of a sublimation till. Ground-probing radar data acquired by the SHARAD instrument onboard the Mars Reconnaissance Orbiter has independently supported the presence of ice within LVF/LDA deposits (Holt et al., 2008; Plaut et al., 2009). Our work further supports the glacial interpretation of Head et al (2006a,b) and through the mapping of the surface lineations on LVF, we find that the deposits surrounding Sinton exhibit large-scale patterns of integrated flow that cover a surface area $>10,000 \text{ km}^2$. Evidence for subsequent episodes of smaller scale glacial activity is also

reported and compared to studies by Levy et al. (2007) and Dickson et al. (2008). Together, the first two chapters support the occurrence of punctuated glacial activity along the dichotomy boundary that occurred during the Amazonian and portions of the Hesperian.

Over the last decade the arrival of high-resolution cameras into Mars orbit (MOC, THEMIS, HRSC, CTX and HiRISE) has permitted the identification of small-scale young landforms that are suggestive of the action of liquid water and ice. For example, the discovery of ~1 km-long gully features that are found within the mid and high-latitudes are interpreted to have been eroded by the recent flow of liquid water (Malin and Edgett, 2000). The existence of gullies and other such water/ice-related features demonstrate the preservation of Late Amazonian climatic changes within the geomorphic record. In order to contribute to the investigation of this most recent period of martian history, we conducted field work in the Antarctic Dry Valleys (ADV). The ADV are a hyperarid, cold polar desert (Marchant and Head, 2007) and have long been considered to be the closest terrestrial analog to Mars in terms of temperature, precipitation and physical setting (e.g. Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005; Baker, 2001). Gullies, morphological similar to those identified on Mars, are present in the ADV (Marchant and Head, 2007). Chapter Three describes the fluvial activity observed in ADV gullies during the 2006-7 field season. Field investigations of ADV gully activity shows that it occurs due to the melting of annual and perennial snowpacks under peak diurnal insolation conditions, and documentation of this process forms the basis of a compelling model of gully formation on Mars. This work supports previous research that has suggested that atmospherically-sourced deposits of snow and ice formed the martian gullies (Lee et al., 2001; Costard et al., 2002; Christensen, 2003; Dickson et al., 2007; Head et al., 2008).

Chapter Four builds upon the previous chapter by combining an area solar radiation model with a high-resolution Digital Terrain Model (DTM) of the ADV study region to explore the range of insolation conditions, and to assess how these may be related to gully-forming conditions. The spatial and temporal variation in solar radiation is compared with the location of

gully systems and our records of gully activity in order to further constrain the gully formation process. This work demonstrates the sensitivity of ADV gully activity to specific insolation conditions and suggests a similar relationship for Mars. This is supported by the strong latitude- and aspect-dependence of Martian gullies revealed in global surveys (Costard et al., 2002; Heldmann and Mellon, 2004; Dickson et al., 2007).

In Chapters Five and Six, we return to Mars to apply the insights gained from our ADV fieldwork. Both chapters concentrate on an unusual, highly modified Noachian crater that contains an annulus of ~1 - 2 km deep, and 10 - 20 km wide valley network systems. The combination of 1) steep slopes ($>20^\circ$), 2) abundant debris (from eroding lava layers that fill the crater floor and thus cap the post-crater fill valleys), and 3) a distinctive topographic configuration that optimizes an unusually wide range of insolation conditions, makes this a rich site for the formation and preservation of ice related, Late Amazonian landforms. Chapter Five, investigates the aspect-dependent range in gully morphology present along the valley walls. Through the combination of 1-D version of the atmospheric Laboratoire de Meteorologie Dynamique/Global Circulation Model (Forget et al., 1999, Costard et al., 2002) with a HRSC DTM, we were able to apply the methodology developed in Chapter four for the ADV, to the crater study area. Recent simulations of the variations of Mars' orbital parameters (obliquity, eccentricity and argument of perihelion) have been extended back to ~20 Ma (Laskar et al., 2004) and provide a temporal framework from which to help constrain models of recent climatic change. Using Laskar et al. (2004) simulations as the input to the climate model, we were able to ascertain that the pole-facing gullies were likely to have been active more recently than the gullies on equator-facing slopes. This result is consistent with the more degraded morphology of the equator-facing gullies and the occurrence of higher numbers of superimposed impact craters relative to the pole-facing gullies.

In Chapter Six we conduct a complete geomorphological study of the crater site in order to establish the local climatic history and make comparisons with similar features reported else

where on Mars (e.g. Milliken et al., 2003). This study resulted in the discovery of Lobate Debris Tongues (LDT), a new class of ice-rich flow features that is superimposed by the gully systems discussed in chapter five. The LDT are similar to the ice-cored lobes that we studied in the ADV. Thus, we argue that the study site underwent a climatic transition from one compatible with excess snowfall and formation of debris-covered glaciers, to one incompatible with excess snowfall and glaciation, but consistent with minor snowmelt and gully formation.

Chapter Seven provides a synthesis of all the previous chapters and places the resulting research within the context of a geological timescale framework. This is compared with the martian literature in order to place limits on our current conception of the nature of the post-Noachian climate. The chapter closes the thesis by outlining outstanding questions that need to be addressed in martian and Antarctic sciences over the upcoming years and suggests further goals for the Mars research program.

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Abbreviations used in the Text

ADV	Antarctic Dry Valleys
CTX	Context Camera (onboard Mars Reconnaissance Orbiter)
CTZ	Costal Thaw Zone of the Antarctic Dry Valleys
DEM	Digital Elevation Model
DTM	Digital Terrain Model
EF	Equator Facing Slopes
GLF	Glacier Like Feature (Afstrom and Hartmann, 2005)
HiRISE	High Resolution Imaging Science Experiment (onboard Mars Reconnaissance Orbiter)
HRSC	High Resolution Stereo Camera (onboard Mars Express)
IMZ	Inland Mixed Zone of the Antarctic Dry Valleys
LDA	Lobate Debris Apron
LDT	Lobate Debris Tongue
LDM	Latitude Dependent Mantle
LIDAR	Light Detecting and Ranging (Used to generate DTMs)
LVF	Lineated Valley Fill
MOC	Mars Orbital Camera (onboard Mars Global Surveyor)
MOLA	Mars Orbital Laser Altimeter (onboard Mars Global Surveyor)
PF	Pole Facing Slopes
RADAR	Radio Detection and Ranging
SF	South Fork of Upper Wright Valley of the Antarctic Dry Valleys
SUZ	Stable Upland Zone of the Antarctic Dry Valleys
THEMIS	Thermal Emission Imaging System (onboard Mars Odyssey)
VFF	Viscous Flow Feature (Millikan et al, 2003)

Acknowledgements

I really do feel very fortunate to have had the opportunity to come to Brown University to work on a Ph.D. It has been a great four years and it is safe to say that I learnt a lot here in the Department of Geological Sciences. Therefore, it is only fair that I take this opportunity to thank the many people who have helped me to cross an ocean and get the most out of almost half a decade in Rhode Island.

Firstly, I need to thank my advisor, **James W. Head**. Jim is not only the reason I have finished this thesis but he is the person who gave me the opportunity to travel to the United States and work on Mars. I have really appreciated his advice and support and it is great to say that I can call my advisor a friend. Being involved with NASA has been a dream of mine for as long as I can remember, and so it has been wonderful to meet Apollo astronauts, be involved in planetary missions and have a beer or two while discussing Mars. Jim has also given me many an opportunity to pursue another one of my passions, travel. The most amazing destination has to be Antarctica and a place I think I have found my calling. What can I say, it's been a real education.

While on the topic of Antarctica, I need to give a thank you to **David Marchant** of Boston University. Dave has led the two field seasons I have been on in Antarctica and I have really learnt a lot from him. Ever since my first visit to the Dry Valleys, Dave has been a coauthor on many projects and I hope to work with him in the future.

Completing my graduate education at Brown University wouldn't have been possible without the advice and support of my prelim and defense committees. Therefore, I would like to take this opportunity to thank **Carle Pieters, Timothy Herbert, Mike Wyatt, Jim Russell** and **Augustin Chicarro** for their valued time.

One of the truly great aspects about my time at Brown has been the friends I have had the opportunity to make and the community spirit of the Geological Sciences Department, which has made working on my Ph.D both enjoyable and memorable. There are a lot of people that I need to

thank for being a source of fun and laughter, so here is a big thank you to: **Wes Patterson, Noah Petro, Carlos Rincon, Caitlin Chazen, Bethany Ehlmann, Brendan Hermalyn, Peter Isaacson, Seth Kadish, Joseph Levy, Mariela Salas-De La Cruz, Jay Dickson, Caleb Fassett, David Shean, Mark Salvatore, Debra Hurwitz, James Jepson, Yannis, Matt Grigsby, Laura Kerber, Sam Schon, and Alish Kress.** Its been emotional.

In addition to the above, it is only fair that I write a few more lines to honour some of the people who have really gone out of their way to help me start and finish this thesis.

Wes Patterson, who really helped me get set up in Providence and has been a source of information ever since. He also has gone out of his way to watch rugby with me and so after living in a country of baseball and American ‘football’ I will always be grateful for that!

Caleb Fassett, who has really taken a lead on ensuring all of us within the research group has had more data than we could ever hope to handle and has been the source of many insightful scientific discussions and a bottomless well of technical insight. Caleb, thank you for all the scripts!

Jay Dickson, a friend, coauthor and finally a gym partner. Thanks for the most fun two guys in a tent on a ‘rock glacier’ on Christmas Day could hope to have.

Laura Kerber. A great office mate and a rich vein of insight. Laura has spent many an hour helping me with research when she should have been doing fluid dynamics, and for that I owe her a big thank you for several of the chapters presented here.

Finally I need to say another big thank you to a group of people who were with me to the very end (in terms of making sure this thesis was finished!). These were **Debra Hurwit, Mark Salvatore, Jay Dickson** and **Brendan Hermalyn**, their encouragement and ability to print out and punch holes in large amounts of paper helped me get through the last ‘push’ of the thesis.

Finally, the person who has helped me get through the best and worst of this four-year adventure in Providence has been **Alida Camp**. Her support, strength, laughter and encouragement has pulled me through and made the Ocean State a real home to me. Thank you

for flooding me with letters while I camped in the freezing Dry Valleys and tea while working at my laptop at home! In addition to the fun we have had together, Alida has taught me a great deal and I am a far richer person for knowing her as a result.

I have been very fortunate in life that I have had a great family behind me who have taught me a massive amount. I wouldn't have made it here if it were not for the hard work of my **Grandparents** who really made the very most out of their lives in the Valleys of Gwent. My **father** has been with me every step of the way and deserves more credit for this thesis than me as a result. His input, advice and friendship in life has been more than any son could ask for. I also need to give a mention to my **Uncle Steve**. Always behind me and a wonderful bloke.

Finally, the person I owe the most too is my **mother**. Sadly she didn't get the opportunity to see me complete this thesis, but the degree to which she encouraged and believed in me in life, I know she never had any doubt that I would finally finish it. The dignity and courage my mum showed to the end will always be an inspiration to me. She is greatly missed.

This Thesis is dedicated to the memory of my mum:

June Morgan

“Ser bore a ddwyreynt yn llu i gyd-ganu gynt..”

Goronwy Owen, Gwaith y Parch 1860

[The stars of morning rose in a host to sing together of old]

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Chapter One

Sinton Crater, Mars: Evidence for Impact Into a Plateau Icefield and Melting to Produce Valley Networks at the Hesperian-Amazonian Boundary.

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Published in current form as:

Morgan, G.A., Head, J.W., 2009. Sinton Crater, Mars: Evidence for Impact Into a Plateau Icefield and Melting to Produce Valley Networks at the Hesperian-Amazonian Boundary. *Icarus*.

10.1016/j.icarus.2009.02.025.

Abstract

The majority of Martian valley networks are found on Noachian-aged terrain and are attributed to be the result of a ‘warm and wet’ climate that prevailed early in Mars’ history. Younger valleys have been identified, though these are largely interpreted to be the result of localized conditions associated with the melting of ice from endogenic heat sources. Sinton crater, a 60 km diameter impact basin in the Deuteronilus Mensae region of the dichotomy boundary, is characterized by small anastomosing valley networks that are located radial to the crater rim. Large scale deposits, interpreted to be the remains of debris covered glaciers, have been identified in the area surrounding Sinton, and our observations have revealed the occurrence of an ice rich fill deposit within the crater itself. We have conducted a detailed investigation into the Sinton valley networks with all the available remote data sets and have dated their formation to the Amazonian/Hesperian boundary. The spatial and temporal association between Sinton crater and the valley networks suggest that the impact was responsible for their formation. We find that the energy provided by an asteroid impact into surficial deposits of snow/ice is sufficient to generate the required volumes of melt water needed for the valley formation. We therefore interpret these valleys to represent a distinct class of martian valley networks. This example demonstrates the potential for impacts to cause the onset of fluvial erosion on Mars. Our results also suggest that periods of glacial activity occurred throughout the Amazonian and into the Hesperian in association with variations in spin orbital parameters.

I. Introduction

Valley networks commonly occur as extensive systems in the Noachian highlands (Carr, 1996) and are often connected by large open-basin lakes (e.g., Fassett and Head, 2008), suggesting widespread fluvial erosion and drainage toward the northern lowlands. Most valley

networks ceased forming at about the Noachian-Hesperian boundary (e.g., Fassett and Head, 2007). The few that are observed to have formed during the Hesperian appear in different settings (e.g., volcanic edifices and plateaus near outflow channel sources) and appear to be due to local environmental conditions (e.g., Fassett and Head; 2005., 2006; Mangold et al., 2008) rather than a major reversion to Noachian climate conditions.

Sinton is located within an isolated plateau along the northern dichotomy boundary at Deuteronilus Mensae (Fig. 1). The majority of the valleys display broad, steep walls and anastomosing patterns around streamlined landforms. Despite extensive surveys of the surfaces of the surrounding plateaus and highlands, no other fluvial erosional features were identified in the vicinity. This suggests that the valley formation was related to the impact that created Sinton crater. Crater counts along the ejecta blanket of Sinton on which the valleys are situated provides a maximum age for the impact event close to the Hesperian/Amazonian boundary.

The study region is also host to extensive deposits of lineated valley fill (LVF) and lobate debris aprons (LDA) which have been interpreted to represent the remains of debris covered glacial systems (Lucchitta, 1984; Head et al, 2006a, b; Morgan et al., 2008). Evidence exists to suggest that multiple episodes of LVF emplacement have occurred in the study area (Morgan et al., 2009). We therefore explore the possibility that the impact occurred at a time when the plateau surface was covered in ice. Such a scenario represents a special case of the impact induced valley formation model first proposed by Brackenridge et al. (1985), and thus the Sinton valleys should appear to represent a new class of valley network associated with local environmental conditions. If our interpretation is correct it implies that snow and ice deposition along the dichotomy boundary was extensive during the Late Hesperian-Early Amazonian. This suggests that Mars has undergone significant ‘glacial’ periods most likely associated with obliquity variations through large portions of its history. The associated climatic variations may have been responsible for significant modification of the dichotomy boundary. In the following

sections we discuss the morphology of the valleys, review the evidence for their formation and develop criteria to recognize similar features elsewhere on Mars.

2. Geology of the Sinton Region

The study area is located within the Deuteronilus Mensae region of the northern dichotomy boundary (Fig. 1). This portion of the boundary forms an escarpment separating the highlands from the northern plains. The region is characterized by fretted valleys, which divide the most northern reaches of the highlands into plateaus and mesas that become progressively smaller to the north. These remnant fragments of the highlands are considered to be typical of the material that comprise Noachis Terra to the south and are interpreted to be composed of both sediments and volcanic material that has been mixed and reworked by impacts (Tanaka et al., 2005). The fretted valleys within the study region and throughout the northern dichotomy boundary are filled with Lineated Valley Fill (LVF) and the walls of the plateau and mesas are flanked with Lobate Debris Aprons (LDA). Much debate has surrounded the origin of the LVF/LDA deposits, though most authors agree that ice was involved in their formation (Carr and Schaber, 1977; Squyres, 1978; Lucchitta, 1984; Head et al., 2006a, b). Comprehensive examinations of the morphological properties of the deposits have argued that they are the remnants of debris-covered glaciers (e.g. Lucchitta, 1984; Head et al., 2006a,b). Morgan et al. (2009) report on detailed investigations into the LVF/LDA deposits in this region and found evidence for integrated flow patterns interpreted to represent extensive valley glacial deposits, supporting previous work concluding that Amazonian glaciation played a role in the degradation of the fretted valleys (e.g. Head et al., 2006a, b).

2.1 Plateau Surface

The main plateau on which Sinton crater is situated extends for over 300 km in a NW by SE direction and is ~ 100 km at its widest extent (Fig. 2). It is surrounded on all sides by steep slopes (Fig. 2b) that rise over 1 km from the surrounding LDA and LVF. The plateau has been mapped by Tanaka et al (2005) to consist of highland material and represents the same unit as the elevated terrain to the south. From the outer flanks of the plateau, there is a more gradual rise to the center, which culminates in a peak elevation of ~700 m above martian datum. This topographic rise represents the western rim of an ancient highly degraded 350 km Ismeniae Fossae impact basin. The largest crater on the plateau is the well-preserved Sinton crater which is flanked to the east by a ~30 km depression (presumably a relict impact crater) that is degraded to the point that it is only clearly identifiable in MOLA topographic data (Fig. 2). The resulting topography between these two impact features is a large escarpment tens of kilometers in length that runs in a SE direction from Sinton crater. Alcoves, several kilometers across, have been eroded into the flanks of the plateau, and extend for up to tens of kilometer into the plateau center, forming elongated valleys. The majority of alcoves are located along the south outer flanks, though some are found along cliffs facing all orientations.

2.2 LVF/LDA in the Study Region

Large-scale deposits of LVF/LDA are abundant throughout the region surrounding the main plateau. The large ~ 20 km wide valley to the south of the plateau is completely filled with LVF, as are the valleys to the east (Fig. 2). LDA deposits predominately emanate from the northern flanks of the plateau and are also located around all of the surrounding isolated mesas (Fig. 2). Detailed studies of the region by Morgan et al. (2009) have demonstrated that both LDA and LVF systems are fully integrated and likely represent different manifestations of the same material. Small lobes of LVF (~3 km wide) which emanate from within the alcoves merge with,

and feed the main deposits of LVF/LDA (Morgan et al., 2009). Lineations of flow can be traced across the surface of the LVF from the alcoves to where the deposits open up as LDA onto the surrounding plains. This demonstrates that the deposits of LVF/LDA represent large scale integrated systems that extend over areas of 30,000 km² within the study area (Head et al., 2006b; Morgan et al., 2009). Integrated systems of this scale are consistent with terrestrial glacial landsystems, which originate within sheltered cirques and open up into large-scale valley glaciers. Head et al (2006a, b) have argued that surface lag deposits have preserved underlying ice from sublimation in a manner analogous to the sublimation tills found on the surface of debris-covered glaciers in the McMurdo Dry Valleys of Antarctica (Marchant and Head, 2007).

Martian general circulation models indicate that the formation of glaciers could have been initiated during previous climatic regimes associated with the cyclic transition from higher (35°) to lower (present) obliquity values (Madeleine et al., 2007). Under these conditions there is a redistribution of volatiles from large-scale glacial systems on the flanks of the Tharsis Mons (e.g. Head and Marchat, 2003; Shean et al., 2007) to the dichotomy boundary where it is deposited as snow.

Some source lobes of LVF within the study area (Morgan et al., 2009) emerge from alcoves and are superimposed on, rather than integrated with, the main trunk LVF/LDA deposits. These occurrences argue for a subsequent period of ice emplacement to the one that deposited the main trunk LVF/LDA deposits. This supports previous observations of similar features to the east (Levy et al., 2007, Dickson et al., 2008a), suggesting that multiple episodes of glacial activity have occurred along the study region and the northern dichotomy boundary as a whole.

2.3 Sinton Crater

The classification of craters on Mars is made difficult by the diversity of primary impact morphologies relative to the other terrestrial bodies (Strom et al., 1992); nevertheless, Sinton is

clearly a typical complex crater with a diameter of 63 km, a central peak and internal wall terraces. The complex topography of the main plateau results in the crest of the crater rim having an uneven elevation. The highest sections are to the southeast and exhibit an elevation of ~660 m, which is ~1990 m above the lowest to the north (Fig. 2). Such a setting has undoubtedly had an effect on the modification stages of the crater development, causing the wall terraces to be most developed along the southern flanks of the crater.

The interior of the crater (Fig. 3) as well as the surface of some of the terraces are covered by a material that appears largely smooth at HRSC resolution, implying that some form of deposition has occurred in the crater since the formation of Sinton (Fig. 4). Although largely smooth, there are some interesting features within the crater fill material. MOC images reveal a complex dissected surface texture, comprised of a range of morphologies from small pits and buttes, tens of meters across, to more gradual undulating surfaces (Fig. 4a). To the south of the central uplift there is a region of pitted terrain (Fig. 3) visible on coarser-resolution images that consists of closely spaced depressions of the order of hundreds of meters across (Fig. 4b). Such crater fill is common within craters of this size (30 – 100 km) within the mid-latitudes (Crown and Bleamaster, 2007). The morphology is similar to the nature of surfaces of the LVF and LDA within the study area and across the dichotomy boundary as a whole. Such surface textures are interpreted to represent the sublimation of ice from within the LVF/LDA (Carr, 2001, Malin and Edgett, 2001, Mangold, 2003, Levy et al., 2009). Gradual small-scale mass wasting along fractures within the surface of LVF deposits has been envisioned as the process of formation; at temperatures exceeding the frost point, interstitial ice in contact with the atmosphere sublimates, destabilizing solid grains of debris and causing the lateral regression of the fracture to produce a pit (Mangold, 2003). Although the interpretation of LVF/LDA hypothesized by these authors is still debated, the presence of potential sublimation features points to the important role that ice appears to play in the development of the landforms. The ice-related interpretation of such features is also supported by the occurrence of degraded craters with a unique morphology that

are found exclusively on the surface of LDA/LVF (Mangold, 2003). Such craters, called ‘oyster shell’ and ‘ghost’ craters, are present within the surface of the crater fill (Fig. 4c, d) and do not appear to have experienced conventional modes of erosion, but are interpreted to be the result of the loss of volatiles from within the material into which the projectiles impacted (Mangold, 2003), or primary crater forms (ring-mold craters) subsequently degraded (Kress et al, 2008).

Patterns indicative of flow within the crater fill material are evident along the interior crater flanks. This also supports the interpretation that Sinton crater fill is composed of the same material as LVF/LDA. Lobes bounded by concentric raised ridges are present within the fill material occupying crater rim terraces close to the top of crater rim (Fig. 4 e). These are morphologically similar to the lobes of LVF identified within the alcoves that surround the plateau (Morgan et al., 2009) and thus further suggest that the crater fill consists of LVF/LDA material. The lobes are up to 3 km in length, though their size seems to be controlled by the scale of the terrace on which they occur. The presence of LVF/LDA-like material within Sinton crater suggests that the emplacement of ice within the study region is not just restricted to the fretted valleys, but also occurred on the main plateau itself. This is further supported by the occurrence of fill material within other relatively large craters on the plateau.

2.4 Ejecta Deposit and Distribution

The distribution of ejecta associated with Sinton crater is made complex by the fact that the projectile that formed it impacted an isolated plateau that is of a size comparable to the resulting crater. Nevertheless the crater displays an uneven and largely poorly defined ejecta blanket around the remainder of the plateau (Fig 2). To the southeast, there is an apparent radial pattern produced by chains of secondary impacts that form striation-like gouges orientated perpendicular to the crater rim. Such ejecta-related disturbances across the plateau surface are particularly evident at MOC resolution. To the north of the crater on the adjacent lowland plains

there is an accumulation of debris which is likely to have been formed from material ejected during the impact, representing the northern extent of the ejecta deposit (Fig. 2).

The influence of the impact is also present on the highland surfaces to the south (Figs. 2, 5). This is predominantly observed in the form of secondary craters. These are identified by their characteristic morphology, being both shallow and elliptical with their longest axis orientated perpendicular to the crater rim. Despite the extensive distribution of the ejecta to the south, there is no disruption present on the LVF surface within the valley to the south of the main plateau, or in other areas of LVF in the immediate vicinity of the plateau flanks (Fig 5, d). The lack of ejecta and secondaries on the LVF to the south is interpreted to be due to resurfacing subsequent to the impact event.

The most outstanding aspect of the plateau surface surrounding the crater is the occurrence of dense valley networks (Fig. 2). This characteristic makes it truly unique relative to other craters within the region. Figure 5 compares Sinton with three different craters displaying the range of morphologies present in craters of a similar size and freshness across the northern plains, where uneven topography does not have a major affect on the distribution of ejecta. Sinton ejecta shares some similarities with Arandas crater (Fig. 5c); which has a strong radial fabric and the unnamed crater in Figure 5d; which has a ballistic ejecta containing strings of secondaries reminiscent of the secondary chains observed on the surface of the plateau to the south of the study area. The major difference though between Sinton and other craters of a similar size is that the ejecta deposit around Sinton is very poorly defined and far less apparent than the other craters despite similar levels of freshness. This suggests that subsequent to the impact, unusual erosional processes have operated to modify the Sinton crater ejecta deposit.

3. Valley Networks

3.1 Valley Networks on the Plateau Surface

The majority of valley networks are located directly to the south and southeast of Sinton, though some are found as far as 75 km from the crater rim crest (Fig. 2, 5). Many of the valleys have an orientation in a direction perpendicular to the crater rim. The valleys exhibit a range of widths from 30-500 m (the smaller ones being only evident in MOC/HiRISE images), and the longest networks extend > 40 km (Figs. 6-10). The valleys are generally steep-sided and have flat floors (Figs. 7, 9, 10), although it is not clear if the flat floors are primary or are a result of the deposition of material within them. No channels are observed within any of the valleys at any of the available resolutions (Figs. 6-10). Tributary systems are a common occurrence for valleys of all scales and in some instances valleys diverge into several branches that remerge downslope, producing anastomosing patterns (Fig. 10). The source of the valleys is poorly defined; the majority of the valleys either emerge from broad topographic depressions close to the rim of Sinton, or simply originate from plains adjacent to the crater rim crest (Figs. 2, 6). Valleys are observed to disappear into smooth hollows that are located across the plateau and reemerge from the down-slope side (Figs. 2, 7, 11). These depressions occur at a range of scales (of the order of several kilometers) and shapes, and their origin is uncertain, although they may predate or have been formed during the impact of Sinton. The valleys are found in three broad regions around the outer rim of Sinton: to the southeast, directly to the south and to the west (Fig. 2). We investigated each of these areas individually with all available data sets and describe our findings in detail below.

3.1.1 South East of Sinton Crater

The southeastern portion of the plateau displays the longest valleys, extending for ~50

km from the rim of Sinton to the plateau outer flanks (Fig. 6). The plateau in this region is characterized by modest gradients of $\sim 1.5^\circ$ and there are no significant changes in slope associated with the erosional features. The widest valleys are ~ 500 m across, but fluvial erosion features of all scales are present. Despite the meandering displayed by most of the valleys, all of the erosional features are strongly orientated along a northeast – southwest direction, which is perpendicular to the rim of Sinton (Fig. 6 – 9). This is especially evident in the smaller networks, but is also observed in the large valleys. These are aligned in a northeast direction for almost their entire length apart from short, sharp deflections consisting of 90° meanders (Fig. 2, 6 - 9). This suggests that there may be some form of structural control associated with Sinton ejecta that has affected the valleys orientation.

The source of the southeastern valleys is not readily apparent, a situation similar to other valleys across the plateau. Some of the valleys to the east are found to originate close to the drainage divide formed by the summit of the escarpment that runs in a northeastern direction from the main crater rim (Fig. 10). Small valley networks that feed some of the long valley networks are present close to the southern edge of the drainage divide. The northern and eastern side of the escarpment is relatively steeper ($>6^\circ$) than the southern side ($<3^\circ$) and is host to small-scale valleys, ~ 5 km in length. These small-scale networks originate close to the edge of the divide and drain into depressions aligned with the foot of the escarpment (Fig. 10).

The most elevated portions of the valleys are spaced ~ 5 km apart, and are joined by smaller-scale tributaries (Fig. 7). Teardrop-shaped landforms, ~ 200 m long, are present within the wider channels and close to tributary junctions (Fig. 7). Many of the tributaries of the upper portions of the networks join the main trunk valleys as hanging valleys (Fig. 7). Further downslope the valleys coalesce to form complex anastomosing patterns, which open up into broader confluences that are filled with numerous streamlined ‘islands’ (Figs. 8 and 9). Despite the multiple interactions with the smaller scale valleys, the wider and most deeply incised valleys exhibit near constant widths for the majority of their lengths.

Smaller-scale erosional features are resolvable within the valley networks within CTX images. Terraces hundreds of meters wide are present along the outer edge of some of the wider confluences (Fig. 8). Longitudinal groves and etched troughs can also be seen on the floors of channels, although deposition of material within the valleys since their formation may be obscuring other such features elsewhere on the plateau.

3.1.2 South of Sinton Crater

Large numbers of valley networks and associated fluvial features are present along the southern portion of the plateau (Fig. 11). The valleys originate along the outer edge of Sinton and extend to the southern flanks of the plateau. The gradient of this portion of the plateau is slightly steeper, with a value $\sim 2.5^\circ$, than to the northeast, although higher values ($\sim 5^\circ$) are encountered close to the outer rim crest of Sinton where the longer valley networks originate. As was the case to the east, the valleys display anastomosing patterns, and streamlined landforms are prevalent.

In many instances the channels are found to drain into the ~ 5 km alcoves cut into the southern cliffs of the plateau (Figs. 2, 11); this is not only the case for the valleys directly south of Sinton but is also seen to the southeast (Fig. 6) thus raising the possibility that the erosion that generated the channels also played a role in creating or modifying the alcoves (Fig. 11). The occurrence of alcoves without channels elsewhere in the region (Fig. 2), argues that there is no direct relationship between channels and alcoves except that they provide local topographic lows into which the channels drain. There is also no evidence for any form of disturbance on the surface of the LVF within the alcoves associated with the mouths of the valley systems, indicating that the LVF postdates the valleys. High-resolution MOC/HiRISE images of the alcoves reveal that a large number of the channels do not actually cut down into the edge of the alcoves themselves, but stop abruptly 10-100s of meters away from the alcove edge within a region marked by a distinctive transition in surface texture (Figs. 12). The area immediately

surrounding the alcove appears smoother at MOC resolution than that of the surface the valley networks incise, which consist of more hummocky terrain, that likely represents ejecta material from Sinton crater (Fig. 12). This suggests that some erosional process has operated to remove the downstream portion of the valleys and the surrounding surfaces as it seems inconceivable that the valleys would abruptly terminate 100s of meters short of the edge of the alcoves. The removal of material around the alcove argues for post-valley network formation erosion. This is consistent with an episode of alcove enlargement that occurred during a phase of LVF (glacial) activity (Morgan et al., 2009). Indeed the transition between the two units may represent the maximum extent of LVF within the alcove during a period of ice accumulation that occurred subsequent to the formation of the valley networks.

3.1.3 West of Sinton Crater

There are significantly fewer erosional features present directly west of Sinton crater compared to the south and southeastern portions of the plateau. One wide valley network (650 m) does exist and extends due west for 15 km from Sinton, where its terminus is obscured by the ejecta of a fresh 1 km diameter impact crater (Fig. 2). The network exhibits many of the features present in the other valleys, including multiple conduits, steep walls and broad valley floor. A 2 km 'island' has been formed by the divergence and subsequent convergence of the valley networks. The northern most valley is less defined than the one to the south and is comprised of several smaller anastomosing channels (~100 m wide).

3.2 Valley Networks Within Sinton Crater

Small-scale discontinuous valley networks are present on the exposed sections of the internal flanks of the crater (Fig. 13). These valleys are ~300 m at their widest extent and the

longest uninterrupted sections are ~ 10 km in length. The level of incision and density of the valley networks is significantly lower than those on the plateau surface outside of Sinton. The spatial resolution of MOLA data is too low to permit the precise measurement of the depths of the valleys, but estimates from THEMIS images indicates the depths to be of the order of tens of meters. Their distribution is limited to the crater interior walls, which are located below the highest elevated portions of the crater rim to the southeast and northeast (Fig. 3). The valleys are found along the length of these slopes. The widest valleys are located close to the rim crest (Fig. 14). It is difficult to identify the source of these higher-elevation portions of the valleys as the majority emerge from depressions within the upper terraces. No valley networks have been observed at any of the available resolutions on the central peak of Sinton, although they do occur at the same elevations on the surrounding crater wall slopes. The complexity of the valley morphology varies throughout the crater, and appears to be related to slope angle, with the most dendritic drainage configurations consisting of multiple tributaries occurring along the crater terraces, which exhibit slopes close to one degree (Figs. 2, 14). Where the slopes become steeper than this, the valleys show simpler configurations, consisting of increasingly straight channels that are orientated parallel to each other in a downslope direction.

At a number of locations close to the base of the internal slopes of the crater, valley networks open up into elongated alcoves, of the order of a kilometer wide and 5-10 km long. Fan-shaped features several kilometers wide protrude out from the mouths of the alcoves and extend out for ~ 3 km over the plains of the crater floor (Fig. 14). A similar-sized and shaped landform is also present further up the crater interior flanks above the crater floor where there are breaks in slope (#3 in Fig. 14). Valley networks terminate at the upslope portion of the apex of the feature and other networks originate at the downslope margins where there is a prominent increase in slope of $\sim 5^\circ$. At THEMIS resolution, (~ 18 meters/pixel) the surface of all of these features appears to consist of a smooth texture that is similar in nature to the material which occupies the surfaces of the terraces within the internal crater flanks (e.g. #2 in Fig. 13). Other valley networks

also terminate into the similarly smooth textured material comprising the fans, and there are several more examples of accompanying valley networks that originate on the downslope side. In one case the downslope portion of the valley emanates from a theater-shaped depression (Fig. 14).

4. Interpretations of the Valley Networks and Comparisons with other Martian Erosional Features.

4.1 Valley Networks on the Plateau Surface

The availability of high-resolution data sets of the martian surface over the last decade have enabled multiple valley networks and associated erosional features of comparative size to those that surround Sinton to be identified and analyzed in detail. These have included examples present in both Noachian (e.g. Fassett and Head, 2005) and Hesperian (e.g. Ansan and Mangold, 2006; Fassett and Head, 2006) aged terrain, and have been interpreted to have formed over differing periods of time as a result of varying degrees of activity, from prolonged episodes of pluvial activity (e.g. Mangold et al., 2004) to the catastrophic releases of water (Mangold et al., 2008).

A variety of valleys with diverse morphologies have been identified elsewhere on Mars on relatively gentle slopes (similar to those of the surface of Sinton plateau, $< 3^\circ$). Typical Noachian valley networks with sizes comparable to the Sinton external valleys display well-developed meanders and dendritic patterns that consist of multiple tributaries (Carr, 1996). Morphological and morphometric comparisons with terrestrial networks suggest that the martian networks were formed gradually during warmer and wetter climate conditions (Carr, 1996). The identification of Noachian interconnected paleo-crater lakes, requiring sufficient influx of water from feeder valleys to fill the crater so that breaches can occur, further favors clement conditions (Fassett and Head; 2005, 2008). Localized episodes of fluvial activity may have occurred during

the Hesperian in the form of similar dendritic valley systems on 2.8 – 3.4 Gyr old terrain to the west of Echus Chasma (Mangold et al., 2004). The morphology of these valleys contrasts somewhat to the Sinton valleys, which exhibit only limited meandering and are never characterized by more than three clear Strahler's stream orders, four orders less than the valleys west of Echus Chasma.

Valley networks tens of kilometer in length can also form within small-scale outflow channel systems as a result of high discharge, low frequency events. A 60 km long, 50 m deep valley network is present along the outer reaches of an outflow channel south of Nili Fossae in Syrtis Major, interpreted by Mangold et al. (2008) to be the result of subsurface discharge of water due to volcanic activity. This valley shows some similarities with the Sinton valleys (in that limited tributaries feed it), but it differs morphologically in that they begin in theater shaped heads. The scale of incision is also more substantial as the valley is ~30 m deeper than the largest Sinton valley and up to 300 m wider. This is especially significant since the outflow valleys are cut into volcanic plains that are substantially more resistant to erosion than the poorly consolidated ejecta material of the Sinton crater valleys. In addition, the Nili Fossae outflow channel is accompanied by other erosional landforms (e.g. groves and tear drop shaped islands) that are external to the valley itself. Similar features are absent on the Sinton plateau at this scale, perhaps due to a much higher flux of water in Syrtis Major relative to the Sinton valley networks.

The flanks of some of the relatively small volcanoes (180 – 270 km edifice diameter) in Tharsis and Elysium have valleys that extend radially from the outer edge of the summit calderas (Fassett and Head, 2006; 2007). These valleys range in width from 200 m to 2.5 km and have been interpreted to be fluvial in origin. The steeper slopes on the volcanoes (~8°) relative to the Sinton plateau (1.5-3°) are interpreted to have caused the valleys on the volcanoes to consist of sub-parallel, immature drainage patterns (Fassett and Head, 2007). Despite this difference in slope, there are some similarities between the two networks. Both systems display valleys that begin abruptly and have near-constant widths and depths over their entire lengths. This is a

characteristic that has also been noted in fluvial features in the McMurdo Dry Valleys of Antarctica (Fassett and Head, 2006), a region regarded as an applicable terrestrial analog (e.g. Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005).

The general morphology of the Sinton crater valleys is unique relative to other similar-sized valleys previously identified on Mars: they consist of broad flat floors and steep walls; anastomosing drainage patterns; hanging valleys; streamlined islands; and a strong spatial association with an impact crater. The broad expanse of the fluvial features of all scales along the plateau surface (Fig. 2), and especially to the southeast (Fig. 6) indicate that fluvial erosion must have been rather intense, though concentrated on the plateau surface surrounding Sinton crater.

The occurrence of main trunk valleys (more deeply incised into the plateau surface than the smaller-scale fluvial features that surround them) suggests that these larger valleys formed over a longer time period as a result of the initiation of stable drainage configurations concentrating fluid flow along the most efficient drainage pathways (i.e. those with a larger hydraulic radius) (Fig. 15). This would have been accompanied by the abandonment of the smaller (presumably sediment clogged) fluvial pathways, leaving them as relict features and hanging valleys perched on the plateau surface surrounding the main valleys. However, because the smaller valleys have been preserved and thus were not removed by the progressive down cutting and widening of the larger valleys, the fluvial erosion could not have been prolonged sufficiently to enable mature drainage configurations to develop on the plateau surface (Fig. 15). This is further supported by the constant width of the large valleys, which demonstrates their lack of maturity. We therefore interpret the Sinton external valleys to represent a distinct type of martian valley network that was formed and modified by processes potentially differing from those responsible for the erosion of the other types of valleys on Mars.

4.2 Estimation of Discharge.

Estimating the discharge responsible for the erosion of fluvial features from remotely sensed data sets requires several steps, and contains a large degree of uncertainty. Due to the small scale of the valleys, it is difficult to establish their cross-sectional dimensions accurately. Nevertheless, establishing estimates for discharge is important as it allows us to place some constraints on the geologic processes responsible. Komar's (1979) modification of the Manning equation which corrects for martian gravity, has commonly been applied to the calculation of martian fluvial networks:

$$Q = A(g_m s R^{4/3} / g_e n^2)^{1/2} \quad (1)$$

where A is the cross sectional area, R is the hydraulic radius (the ratio of flow cross sectional area to wetted perimeter), g_m and g_e are the values for gravity on Mars and Earth respectively and n is the Manning coefficient. As there is no means to establish the depth of fluid in the valleys, we assume bankfull discharge, which will provide an estimate of maximum discharge. Valley depth was estimated from individual MOLA points and the width was measured directly from the highest resolution images. The valleys were assumed to have had a rectangular cross section. Slope measurements were derived from high resolution HRSC DEMs (~200 m). The Manning coefficient involves additional uncertainty as it is determined empirically for terrestrial channels and cannot be measured directly from the available data. We used the approximation of Wilson et al. (2004) for the coefficient (0.0545) that is more appropriate for Martian conditions. It is unknown whether the fluvial erosion was continuous or if there were several episodes of activity. As discussed in the previous section, there appears to have been a concentration of flow in the larger valley networks at some point after the initiation of the fluvial features (Fig. 15). Therefore, for the purposes of calculating the discharge we have only considered the flow within the largest

valleys from the southeastern, southern and western portions of the plateau. These valleys are also transected by suitable individual MOLA orbital tracks which permitted cross section dimensions to be estimated. The measurements we made and our estimates of discharge are summarized in table 1 below.

The largest valleys surrounding Sinton crater have large estimated peak discharge values of $2.5 \times 10^4 - 1.4 \times 10^5 \text{ m}^3\text{s}^{-1}$ (Table 1), with the upper limit of this range corresponding to the widest network, situated directly west of Sinton (Fig. 13). Although there is uncertainty in the valley cross-sectional profiles and the nature of fluvial erosion under martian conditions (e.g., Fassett and Head, 2005, 2007), the discharge estimates for the Sinton valleys are over two orders of magnitude greater than the 80 km long valley network that enters Jezero crater. The Jezero crater valley was interpreted to have formed over relatively prolonged time periods (minimum time of $\sim 10 - 18$ years) under Late Noachian climatic conditions (e.g., Fassett and Head, 2005). The Sinton discharge estimates are more consistent with valley networks formed as the result of volcanic thermal anomalies releasing significant amounts of water from snow/ice deposits (e.g., Ceraunius Tholus, Fassett and Head, 2007) and groundwater sources (e.g., Syrtis Major Planum outflow channel, Mangold et al., 2008). In a later section, we assess the most probable source of water that would have been capable of supplying such high peak discharges.

4.3 Valley Networks Within Sinton Crater

The drainage configurations of the internal valleys appear to be correlated with slope, with the more dendritic patterns present on low slopes ($<6^\circ$) and the simpler, straighter valleys found on the steeper sections ($10 - 20^\circ$). Such drainage patterns are consistent with other networks on Mars and on the Earth.

The morphology and smooth texture of the fan shaped features which occur where the valleys terminate onto flat surfaces (Fig. 14) is consistent with the material being depositional in

nature and is interpreted to represent sediment that was deposited when the valley networks were active. We interpret the fan shaped features to be alluvial fans deposited by stream flow during this period of activity. The presence of such broad fans suggests that they were built up gradually, perhaps as a result of discrete low frequency events.

The intriguing aspect of these valley networks compared to other martian examples is their: 1) small scale, 2) localized nature, 3) low drainage densities, and 4) discontinuous nature in the downslope direction. Valleys exhibiting discontinuities within their networks are not uncommon on Mars, largely due to the effects of post-formation erosion and deposition over several billion years (e.g., Carr, 1996). In the case of the Sinton interior valleys, their intermittent nature suggests that surface runoff infiltrated into the ground and traveled downslope as throughflow before reemerging at the surface to continue eroding the valley. For example, a valley network can be seen emerging from directly below the outer margins of the fan deposit located along the crater terrace (Fig. 14), supporting this interpretation. Similar relationships between channels and fans are also observed within the small martian gullies first observed by Malin and Edgett (2000). These landforms are slightly smaller (~1.5 km difference in length) than those within Sinton, but exhibit similar morphological characteristics including sinuous channels and depositional fans. The Malin and Edgett (2000) gullies also consist of isolated drainage configurations and are commonly located within crater interiors. Progradational and abandoned channels are prominent on some of the gully fan systems, suggesting that they have experienced multiple episodes of activity. This activity has been attributed to the melting of volatiles during periods of high obliquity (e.g. Christensen, 2003; Dickson et al., 2007), and the gullies are morphologically similar to small-scale fluvial features in the McMurdo Dry Valleys of Antarctica that have been produced by summer meltwater from snow packs (e.g. Morgan et al; 2007, 2008). Observations of the Antarctic gullies found that channel runoff would frequently infiltrate into fans and resurface downslope forming theater shaped channel systems (Morgan et al., 2007; 2008; Marchant and Head, 2007) that are analogous to those within Sinton crater.

Hence, we interpret these features to have been sourced from water infiltrating into the smooth sediments deposited along the terrace, and traveling downslope within the shallow subsurface, before reemerging at the surface as runoff a kilometer downslope.

The differences between the internal and external valleys may be due to the different environments in which they are located (i.e. crater interior compared to plateau surface respectfully). Alternatively it is possible the valleys did not form at the same time and by the same processes that formed the external networks. Internal valley networks have been identified within other impact craters, such as Lyot (Dickson et al, 2008b), but these craters lack the extensive valley networks on their outer flanks, further suggesting that they formed independently of the external Sinton valleys.

5. Age Relations

Dating the landforms in the study area is essential to the understanding of the geological history of the region. The strong spatial correlation between Sinton crater and the valley networks suggest that the impact was related to their formation. The lack of superposition by the crater or any of its ejecta on the valley networks indicates that the valleys postdate its formation, an interpretation that is supported by the occurrence of valley networks within the crater itself. The event that formed Sinton was instantaneous, and thus if the impact can be dated, it would place a time constraint on the development of the region and provide a maximum age for the formation of the valley networks. Successful impact crater size-frequency distribution studies require suitably-sized areas in which to carry out the counts. Due to the absence of a well-defined ejecta deposit surrounding the main crater it is not possible to use this as a unit to constrain the date of the impact by crater-counting techniques. The actual crater interior itself, despite being of sufficient surface area ($\sim 3100 \text{ km}^2$), is also unsuitable due to the presence of widespread post-impact deposits lining its walls and floor (Figs. 2, 3).

Therefore, an alternative method is to directly derive age estimates for the valley networks themselves, an approach which is difficult due to their small surface area relative to the catchments in which they occur. New techniques for directly dating valley networks have been presented by Fassett and Head (2008), utilizing the areas within buffer zones. This method, however, requires larger-scale valley networks than are present in our study area, in order to provide sufficiently large areas in which to count. Nevertheless, due to the high density of networks to the south and southeast of the main crater which completely cover portions of this area (especially to the southwest, Fig. 9) and because the valleys display a uniform level of degradation, it is reasonable to undertake crater counting over this area to provide a minimum age for these sets of networks.

The longitudinal extent of the crater counting area was restricted to include only the area with the highest drainage density. The edge of the summit of the plateau escarpment served as the boundary to the south as did the rim of Sinton crater to the north. This resulted in a crater counting area to cover a region of the plateau of $\sim 3201.5 \text{ km}^2$ (Fig. 16). During the counting process any craters found to be intercepted by valley networks or associated erosional landforms were not counted. The results of the crater counting (Fig. 17) revealed that for craters $> 0.75 \text{ km}$, the size frequency distribution curve fits the isochrons most closely along the Hesperian/Amazonian boundary ($\sim 3 \text{ Gyr}$ using isochrons defined by Hartmann, 2005), thus indicating that the minimum age for the formation of the valleys is within the Late Hesperian.

The crater size-frequency plot also revealed that the distribution curve rolls over at lower crater diameters, causing the data points corresponding to the smallest craters ($< 0.375 \text{ km}$) to cross the isochrons (Fig. 17a). This is interpreted to be due to the occurrence and continual operation of degradation and resurfacing processes that occurred subsequent to the initial emplacement, causing the preferential obliteration of smaller craters. This could be due to the result of aeolian activity and other such small-scale erosional/depositional activity expected to have affected the martian surface over the last 3 billion years.

To further constrain the age of the valley networks, a second area to the west of Sinton crater was also counted (Fig. 15). Although the drainage density is significantly lower than that of the southern area, ejecta from Sinton and valley networks are present on this portion of the plateau (Fig. 13). Therefore, this part of the plateau should be of a similar age or younger than the region to the south. The crater counting area was defined to within one Sinton crater diameter to the west, corresponding to the extent to which the impact would have deposited the greatest volume of ejecta, and provided a crater counting surface area of $\sim 2030 \text{ km}^2$ (Fig. 16). The size-frequency distribution of the counts within this area are consistent with that of the counts to the south, and also provided a best fit along the Hesperian/Amazonian isochron (Fig. 17).

The internal deposits within Sinton crater were dated to establish the minimum age of emplacement for the most elevated deposits of LVF/LDA material in the study region (the most elevated areas of the fill deposits are located along the upper terraces of the crater and are $\sim 545 \text{ m}$ above the highest sections of the valley LVF that surround the plateau). The geological map (Fig. 3) shows the extent of the count area as the crater fill was treated as a single unit (including the ‘pitted’ fill) and formed a counting area of $\sim 1113 \text{ km}^2$. Only craters smaller than 250 m were found to be present on the surface of the crater fill. Nevertheless the crater size-frequency distribution plot shows a best fit to the right of the 100 Myr isochron consistent with an age of $> 100 \text{ Ma} - 500 \text{ Ma}$ (Fig. 18a). This is consistent with the age estimates provided by crater counts on the LVF deposit in the valley to the south of the main plateau by Morgan et al (In Review) (Fig. 18b) as well as other counts conducted on other LVF/LDA deposits across the northern dichotomy boundary (e.g., Mangold, 2003; Levy et al, 2007). This suggests that crater fill emplacement within Sinton occurred at the same time that LVF/LDA emplacement was occurring across dichotomy boundary during the Late Amazonian. Indicating that if the climate models are correct, the redistribution of volatiles from the poles to the lower latitudes during high obliquity resulted in the deposition of snow along the dichotomy boundary, causing widespread ice rich accumulation, which affected all elevations across the boundary

6. Water Sources and Mechanisms of Fluvial Erosion.

Valley networks and associated erosional features have been identified on Mars since the initial Mariner missions of the late 1960s early 1970s (Masursky, 1973) and their origin has been a source of debate ever since. These valley networks have been dated as having formed primarily during the Late Noachian (Fasset and Head, 2007). Due to the implied involvement of surface runoff they have been cited as important morphological indices of the martian paleoclimate, suggesting that early warm and wet conditions once dominated the climatic regime of the planet (Masursky, 1973). At times later than the Noachian/Hesperian boundary, the general consensus is that Mars became cold and dry (e.g. Carr, 1996). Large-scale outflow channels attributed to the catastrophic release of ground water have been the most dominant form of fluvial erosion since this early phase of wetter conditions. However, early studies by Gulick and Baker (1989, 1990) and Gulick (2001), showed the presence and environments of some younger valley networks. Recent research using newer high resolution spacecraft datasets have also discovered younger valley networks across the martian surface; a prominent example being the Hesperian-aged networks documented by Mangold et al. (2004) around the Valles Marineris area. As previously discussed, these are of a comparative scale to those in the Sinton crater, but are eroded into near flat surfaces and are found to exhibit high drainage densities and highly ordered dendritic tributary systems. Mangold et al (2004) attributed such morphological characteristics to the work of precipitation-fed runoff, suggesting either a transitional climate from the Noachian or episodic periods of wetter conditions. However, there are other means by which valley networks can form that do not rely on the prevalence of conditions warmer than the current climate (Mangold et al, 2004). These include: groundwater sapping, water-lubricated debris flows and hydrothermal activity. Therefore, detailed analysis was carried on the Sinton crater valley networks and associated erosional features in order to identify the most suitable source of water and derive a model of formation that is consistent with the geological evidence.

6.1 The Case for Groundwater.

A groundwater source that excludes the necessity for an atmospherically-derived water source is inconsistent with both the topographic setting of the valleys and their morphology. The isolated plateau on which the valleys are situated is in excess of one kilometer above the surrounding terrain, making it unlikely that its surface could be internally connected to any large-scale regional groundwater sources that have been proposed to exist beneath the cryosphere (Clifford, 1993; Clifford and Parker, 2001). Perched aquifers have been suggested to exist on Mars to explain the existence of ~ 1 km long gullies (Malin and Edgett, 2000), although in the absence of a significant means of recharge, such localized aquifers within the plateau would be insufficient to supply the > 50 km long Sinton valleys. Moreover, the occurrence of valley heads at the crest of steep ridges running along the plateau (Fig. 10) is also inconsistent with an internal water source eroding the valleys. The lack of theater-shaped heads, which would be expected if sapping had occurred (Laity and Malin, 1985) or the absence of chaotic terrain indicative of the rapid release of aquifers (e.g. Baker and Milton, 1974) further argues against an internal water source.

6.2 The Case for Rainfall.

The concentration of valley networks on the main plateau, and their spatial association with Sinton Crater itself (Fig. 2), is not typical of widespread pluvial activity. Estimates of discharge are two orders of magnitude higher for the Sinton valley networks (Table 1) compared to similarly scaled Noachian valleys considered to be formed by prolonged fluvial activity (Fassett and Head, 2005). The study of precipitation-fed terrestrial valley networks is based on the work of Horton (1954) who derived a system of quantification based on morphometric

parameters, which has since been applied to martian examples (e.g. Mangold, 2004). Fluvial geometry has been found empirically to be dependent on a range of factors but is critically related to slope. On Earth, dendritic patterns are only observed on relatively flat surfaces, below angles of $\sim 1^\circ$. Above this value, valleys become increasingly more sub-parallel. As a result of the lower surface gravity on Mars such a value would be different than its terrestrial counterpart. However, in the absence of experimental data or observations of active valley networks, it is not possible to derive a satisfactory value with confidence. Therefore, at values close to critical slope angle one has to be cautious when attempting to make direct morphometric comparisons (i.e. such as bifurcation ratios) with terrestrial precipitation-fed valley networks in terms of the origin of the valley geometry. However, there are other aspects of the morphology, which can be utilized to assess a precipitation origin. Regardless of the angle of slope, precipitation-fed valleys will tend to converge to larger, higher order valleys in the downslope direction. The occurrence of anastomosing drainage patterns and broad, flat valleys in our study region is not consistent with this prediction.

6.3 The Case for the Melting of Snow/Ice Deposits.

Atmospherically derived snow and ice deposits are an alternative source of water, providing that the deposits could be heated sufficiently to generate the meltwater required for erosion. Such a scenario is also consistent with the occurrence of ice-rich remnant deposits that have been found at a range of elevations both below and above the average plateau surface (Head et al., 2006b; Morgan et al., 2009; Fig 2, 3) and would also explain the immature drainage configurations of the valleys (Fig. 15). Volcanically heated snow and ice deposits have been interpreted to release sufficient water volumes to generate discharges comparable to 10^4 - 10^5 m^3/s^{-1} estimated for Sinton (Fassett and Head, 2006, 2007). The thermal anomaly associated with the impact event that formed Sinton Crater would have provided an alternative but substantial

energy source, capable of melting surface ice deposits, if they had existed on the plateau prior to the impact. This would also explain the close spatial association between the valleys and Sinton crater ejecta. Hence, interaction between the energy of impact and residual ice deposits presents a mechanism for inducing surface runoff across the plateau surface. Fluidized ejecta around craters on Mars (Fig. 5b) have also been interpreted to result from impact/ice interactions, through the remobilization of ice within the regolith (e.g. Squires, 1989). These features however, do not show evidence for fluvial erosion. Hence, we argue that surfacial ice deposits were required to generate the Sinton valleys.

A period of Late Amazonian glaciation is well established through the extensive research that has been carried out on LVF/LDA deposits across the northern dichotomy (e.g., Head et al., 2006a, b). At least two periods of glaciation appear to have occurred through the redistribution of volatiles from the poles to lower latitudes during high obliquity (Madeleine et al; 2007). Mars is likely to have experienced large obliquity variations throughout its history (Laskar et al; 2004). Therefore, is conceivable that ice deposition also occurred during earlier periods, including the Hesperian.

Hydrothermal systems can be established whenever conductive heat transport is initiated by the coexistence of a fluid phase (usually water) with a heat source (Farmer, 1996). Impact cratering processes provide significant heat sources that can support hydrothermal systems on the Earth and Mars (Newsom, 1980). Detailed studies of the 23 km Haughton impact structure in Devon Island, Arctic Canada (a well established and utilized Mars analog) have documented evidence for post-impact hydrothermal alteration in concentric fault systems that surround the crater rim (Osinski, 2005a, b). Mineralogical analysis of the alteration indicates that temperatures reached at least 200°C, and that the fault systems provided pathways for hot fluids, potentially for several kilometers away from the heat source. Brakenridge et al. (1985) argued that analogous hydrothermal systems may have been widespread early in Mars history, during the period of heavy bombardment, and may have been responsible for Noachian valley networks, instead of a

warmer, wetter climate period. Brakenridge et al. (1985) further hypothesized that ground ice was exploited by impact hydrothermal systems, producing valley networks through headward sapping and down valley fluid flow. This model has been challenged recently by the dating of Noachian valley networks; the age of most of the valley networks is Late Noachian/Early Hesperian, ~ 0.3 Gyr after the late heavy bombardment (Fassett and Head, 2008). This separation in time is too great to maintain active hydrothermal systems, even if one takes into consideration the several million-year estimates for the lifetime of hydrothermal systems within large scale terrestrial impact basins (e.g. the Chicxulub crater, Mexico; Abramov and Kring, 2006).

In light of the evidence for: 1) high discharge rates, 2) immature anastomosing drainage configurations, and 3) extensive fluvial features in close proximity to a large crater in a region interpreted to have experienced significant glacial activity, we propose that the valley networks originated from the release of water due to the deposition of hot ejecta over snow/ice deposits present on the plateau during the impact event. Although the melting of near-surface ground ice (which presently is estimated to begin at a depth of >1 m deep at Sinton's latitude; Jakosky and Haberle (1992)) may also have contributed water to the fluvial activity, the large volumes required to produce the $10^4 - 10^5 \text{ m}^3 \text{ s}^{-1}$ estimated discharges and the lack of similar fluvial features around other craters of comparative size, argues that ground ice alone is insufficient. The release of water from ground ice due to persistent hydrothermal systems could potentially have supplied the valley networks for an extended period after the initial melt of ice. However, if hydrothermal activity similar to that outlined in Brakenridge et al. (1985) did persist after the melting of the initial external ice deposits, it was not sufficient to enable the valleys to develop into mature networks (Fig. 15).

7. Model

On the basis of the analysis of the geology, geomorphology, age relations and topography of the Sinton crater area, we interpret the valley networks to have formed by the transfer of heat

associated with the ejecta of the Sinton impact to pre-existing surficial snow and ice deposits on the plateau. We hypothesize that enough impact kinetic energy was transferred to the ejecta in the form of heat so that the burial of snow and ice in hot ejecta was sufficient to produce melting and drainage, creating the valleys. Here we present a conceptual model that describes the main stage of valley formation developed from our observations and models of the cratering process (Fig. 19).

7.1 Pre-Impact

Prior to the impact, we hypothesize that the dichotomy boundary was experiencing an active ‘glacial period’ in which ice-rich deposits were present across the plateau (Fig. 19a). Such a scenario is supported by the current occurrence of ice-rich and ice-related deposits throughout the study region in the form of: 1) LVF/LDA deposits within the fretted valleys and plains (Fig. 2), 2) ice-related crater fill within Sinton (Fig. 3), and 3) a mantling deposit on the plateau surface. The occurrence of mantling deposits have been interpreted to be partly responsible for the paucity of small craters within the crater count plots (Fig. 17). These relatively recent ice-rich deposits formed during the late Amazonian in conjunction with periods of relatively higher obliquity (Head et al., 2006a; 2006b; Mustard et al., 2001). The accumulation of significant quantities of snow and ice at northern mid-latitudes during certain portions of the obliquity history of Mars has been modeled by Madeleine et al. (2007). They find that at obliquities of 35° and moderate atmospheric dust opacities, widespread precipitation of snow occurs over the dichotomy boundary and persists throughout the year, ultimately producing plateau and valley glacial deposits (Madeleine et al, 2007). Modeling of the obliquity history of Mars shows that these obliquity conditions are not unusual and that this condition is likely to have recurred frequently in the past history of Mars (Laskar, 2004). Furthermore, recent evidence suggests that the current LDA/LVF deposits represent only a remnant of a much larger and more extensive

plateau glaciation that occurred in the late Amazonian (e.g. Dickson et al., 2008a). Thus, there is a high likelihood that extensive ice and snow deposits occurred on the Sinton plateau at different times in the past history of Mars.

7.2 Impact

As the projectile made contact with the surface, energy was transferred into the plateau target material as a combination of shock waves and associated rarefaction waves, causing the formation of a transient cavity and emplacement of ballistic ejecta (Fig. 18b). Estimates of the energy released by the impact can be made by calculating the kinetic energy of a typical impacting projectile by estimating its mass (m) and velocity (v):

$$K.E. = \frac{1}{2}mv^2 \quad (2)$$

Assuming a mass based on chondritic densities, a diameter approximately one tenth of the crater diameter, and an average velocity of Mars crossing-asteroids ($\sim 10 \text{ ms}^{-1}$) (Ivanov, 2001), the impact would have provided substantial energy, of the order of 10^{19} kJ (Table. 2). This value would be an order of magnitude higher if Sinton was the result of a higher velocity, long-period comet impact (40 km s^{-1} ; Steel, 1998).

During the excavation stage (Fig. 19c) ice deposits would have been blanketed in hot ejecta, which would have contributed to melting the ice, causing water to accumulate at the ejecta-ice interface. The proportion of the initial kinetic energy that is converted to thermal energy is poorly constrained for large impacts. One approach for testing whether the energy from Sinton would have been sufficient to generate the valley networks is to calculate the proportion of the original kinetic energy required to generate the volume of meltwater necessary to carve the channels. The mass of ice melted by a given amount of energy is:

$$M_{ice} = E/L \quad (3)$$

where E is the energy available to the melting of ice, and L is its latent heat of fusion (for which we use 335 kJ/kg). Eq. (3) can be used to calculate the mass of ice (and from which the water volume can be calculated) that can be produced for a range of thermal energies supplied to the plateau as a proportion of the original projectile kinetic energy (Table 2). Table 2 demonstrates that only a small proportion of the initial projectile energy is required to generate substantial amounts of melt. Only 5% of the initial energy ($\sim 10^{17}$ kJ) would need to be utilized by the melting of ice to provide a value comparable with the estimates of the volcanic energy available for snowmelt production and Hesperian valley formation attained by Fassett and Head (2006) for the Hecates Tholus volcano. If we only consider the area of the plateau with valley networks (assuming that any ice in the region where the transient crater formed would have been vaporized; this gives an area of $5 \times 10^9 \text{ m}^2$), 5 – 10% of the initial energy would be sufficient to melt an ice deposit 500 m – 1 km thick (Table. 2). This is comparable to the current thickness of dichotomy boundary LVF/LDA deposits (Li et al., 2005).

The sporadic release of water associated with the melting would account for the formation of multiple small anastomosing channels that have no apparent source (Fig. 6-9, 15a). As the flow developed through the continual release of meltwater, runoff would have been concentrated in the most efficient distribution systems as the other channels become clogged with sediment derived from the erosion of the ejecta. This could concentrate erosion and lead to the development of the observed broad-flat channels (Fig. 15b). The broad depressions that connect the valleys (Fig. 10) may represent where the melting of underling ice was concentrated and caused the collapse and further erosion of overlying ejecta material.

During the outflow of water, portions of the ejected material would be transported downslope by the meltwater and preferentially deposited across the plateau surface and into the

surrounding valleys. This could help to explain the highly modified ejecta deposit surrounding Sinton crater compared to the more distinctive ejecta deposits surrounding other similar-sized craters (Fig 5). During the valley formation, water flow may have been controlled by the micro-topography of the original ejecta patterns (such as the radial fabric patterns produced by ballistic secondary impacts). This is consistent with the larger channels to the southeast of the crater exhibiting an orientation perpendicular to the crater rim for almost their entire length, apart from short, sharp local deflections (Fig. 6).

The lack of more mature drainage configurations and broader valley networks suggests that the erosional process was not prolonged and that runoff ceased after the ejecta had cooled or the ice/snow source of meltwater had been depleted. Nevertheless some hot springs may have developed at the location of the extensional faults. These can act as pathways for hot fluids and steam generated at hydrothermal systems associated with the center of the thermal anomaly under the floor of the crater (e.g., Osinski et al, 2005b).

Based on the discharge estimates calculated in section 4.2, the volume of ice that could be melted by the impact in which 5 – 10% of the initial energy was partitioned for this purpose, would have supplied the 15 largest valley networks for ~ 80 - 150 days. The values of discharge are for peak bankfull conditions and so the duration of flow would likely have been significantly longer, by as much as ten times the value predicted above if we consider the uncertainties surrounding the cross sectional dimensions and surface roughness. Estimates from large scale terrestrial impacts (such as the one that formed the Chicxulub crater) indicate that as much as 50% of the projectile original kinetic energy was converted into heat (e.g. Ryder et al., 1996). Therefore, although there are uncertainties in both the discharge and energy estimates, the calculation does at least demonstrate the physical plausibility of the model and suggests that the fluvial activity was likely to have been short-lived relative to the mature Noachian valleys.

The uneven distribution of the valley networks could represent a combination of the location of ice on the plateau surface prior to the impact, the dispersal of ejecta (related to the

azimuth and zenith angles of the projectile trajectory) and the underlying plateau topography. In our model we argue that the internal valley networks formed after the formation of external valley networks and thus are related to different times and processes. Their morphologic similarities with smaller scale gully systems may suggest that they have formed due by similar processes (e.g. Christensen, 2003., Dickson et al., 2007).

8. Summary and Conclusions

Previous analysis of the dichotomy boundary region have shown an array of evidence for significant glacial activity occurring in the form of LDA and LVF throughout the region, and specifically in the area surrounding Sinton impact crater (Morgan et al., 2009). Large scale integrated LVF and LDA systems surrounding the main plateau have been identified and mapped and found to originate from within multiple alcoves that are cut into the plateau flanks (Head et al., 2006b; Morgan et al., 2009). Ice related deposits are also prevalent across the surface of the plateau within topographic lows, such as impact craters. Sinton crater itself is host to a large-scale unit of crater fill, exhibiting a range of ice and water related flow features (including kilometer-scale lobes) that have been identified within other similarly sized craters in the mid-latitudes. Our data therefore suggests that significant amounts of ice have been deposited within the study area on the surface of the plateau in the past.

The spatial relationship between the valleys and the main crater, and the apparent control its ejecta has imposed on the orientation of valley formation, suggest that the two are related. The thermal anomaly associated with the impact provides a means of melting ice deposits that were present on the plateau at the time of impact. Hence an ice-ejecta interaction forms the basis of our valley network formation model; the release of water is initiated by the melting of ice from the deposition of hot ejecta deposits over its surface. Such a mechanism provides a means of generating intermediate scale fluvial features (i.e. hundreds of meters wide by tens of kilometers

long) in the absence of a climatic regime favorable for pluvial activity.

Previous work has suggested that there was a major period of climate change in the Hesperian, transitioning from a potentially "warm and wet" Noachian, to a "cold and dry" Amazonian (e.g., summarized in Carr, 1996). Our observations of Sinton crater provide evidence that a significant cover of snow and ice existed on the dichotomy boundary plateau during at least part of the Hesperian (see Fig. 20). Our results also demonstrated how the interaction of ice with localized heat sources can generate intermediate-scaled fluvial landforms and thus is consistent with the interpretation that other such thermal anomalies (e.g. such as those associated with volcanic edifices; Fassett and Head, 2007) are capable of generating runoff through the melting of atmospherically deposited snow/ice. Glacial activity could therefore have accounted for one of the dominant erosional processes operating along the dichotomy boundary since its formation.

Future work will concentrate on answering two outstanding questions: 1) What was the volume and spatial extent of ice during a 'glacial' period? 2) How many glacial cycles were there and for how long did the activity last? One approach that could be implemented to begin to answer these questions is to investigate other valley networks in association with large craters, as the identification of other impact/ice-generated valley networks provides both temporal and spatial constraints on the deposition of ice during past martian history.

Acknowledgments

We gratefully acknowledge financial support from the NASA Mars Data Analysis program (NNX07AN95G), the NASA US participation in the Mars Express High-Resolution Stereo Camera (HRSC) (JPL1237163), and the NASA Applied Information Systems Research Program (NNG05GA61G). Thanks are also extended to Caleb Fassett, James Dickson and Wes Patterson for their technical assistance in data analysis and contribution to scientific discussions during preparation of the manuscript.

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Tables

Valley System	Width (m)	Depth (m)	Slope (°)	Q (m ³ s ⁻¹)
Sinton Southeast	320	20	1.5	8 x 10 ⁴
Sinton South	234	12	2.8	2.5 x 10 ⁴
Sinton West	480	50	2	1.4 x 10 ⁵
Valley entering Jezero Crater, Nili Fossae (<i>Fassett and Head, 2005</i>)	170 - 400	-	-	500 – 900
Ceraunius Tholus Valleys (<i>Fassett and Head, 2007</i>)	250 - 700	20 - 60	5 - 8	4 x 10 ⁴ – 6 x 10 ⁵
Syrtis Major Planum Outflow Channel. (<i>Mangold et al., 2008</i>).	500	50	0.7	5 x 10 ⁵

Table 1. Estimates of channel-forming discharge for the largest valley within each region of the plateau (using Komar (1979) modification of the Manning equation, and Wilson et al. (2004) estimate of the Manning coefficient for martian channels) compared to estimates for similar-sized martian valleys.

Percentage of initial projectile kinetic energy	Ice thickness that could be melted (m)	Volume of water (km ³)	Duration of flow within the valleys (Days)
1	109	500	16
3	328	1500	47
5	547	2500	78
10	1090	5000	157
20	2188	10,000	314

Table 2. Energy balance calculations that give the volume and duration of flow for a given percentage of initial projectile kinetic energy supplied to the melting of ice.

Figure Captions

Fig. 1. Context MOLA topographic map of the northern dichotomy boundary at Deuteronilus-Protonilus Mensae; the location of the study region is highlighted by the red box. Note the gradual decrease in topography from the southern highlands into the northern lowlands. The boundary area is characterized by the occurrence of numerous mesas which become increasingly smaller to the north.

Fig. 2. a) Sinton Crater and vicinity. The map highlights the extent of the LVF and LDA systems within the region (in valleys with the outer boundary indicated by the dashed white line), which completely surround the plateau on which Sinton is located (portion of HRSC image 1589, overlain by MOLA topographic data). b) Slope map. The fretted terrain within the study area consists of relatively smooth-surfaced plateaus and mesas surrounded by steep flanks exhibiting slopes of over 23° (derived from the DTM). Smooth crater-fill deposits line the floor of Sinton crater and the terraces along its interior flanks. The LVF/LDA deposits surrounding the main plateau exhibit almost entirely horizontal surfaces, which consist of slopes less than 0.5° (dark blue). Note the presence of a steep topographic ridge on the surface of the main plateau that extends to the southeast from Sinton crater. Slope map overlain on a portion of HRSC image 1589. c) Map of the distribution of: i) Valley networks. Note that the valleys are located around Sinton crater, and the majority originate close to crater rim. The regions depicted in white represent topographic hollows within the plateau. Note that valleys originate close to the NE topographic ridge along the plateau surface that can be seen in b). ii) Secondary craters and ejecta material (highlighted in yellow) interpreted to have been formed by the impact that created Sinton crater. HRSC image 1589. d) 300 m contour lines overlain on a Viking image mosaic. Note that the LVF/LDA surrounding the main plateau can be seen clearly in the image (compare with (a)).

Fig. 3. Sinton crater and accompanying geological map of the major units. a) Valley networks emanate from the crater terraces and crater fill deposits. Crater fill deposits appear largely smooth at HRSC resolution (~ 8 m/pixel) and cover the crater floor and terraces. Boxes represent the location of images in Fig. 4. HRSC image 1589. b) Geological sketch map of Sinton crater, showing the four main geomorphic units (see legend) and the distribution of internal valley networks. North is at the top of the image. The map emphasizes the extent of the crater fill deposits which cover a total surface area >1200 km². Small-scale valley networks are present throughout the internal rim of the crater, and are most developed along the wider southern flanks.

Fig. 4. Surface morphologies of Sinton crater fill deposits. a) Comparison between the surface texture of the crater fill material along the floor of Sinton crater (i) and the LVF deposits along the floor of the southern valley system at MOC resolution. The similarity between the two surface morphologies (consisting of knobs and pits) suggests that the deposits share a similar origin. The pitted texture is interpreted to be caused by the action of sublimation (e.g., Mangold, 2003, Levy et al., 2009) and thus suggests that the deposits once contained a significant amount of ice. MOC images: M0300514 (a) R0700879 (b). b) Image of the 100 meter scale pits in the surface of the crater fill south of the central peak. HRSC image 1589. c) Image of a ‘ghost’ crater in the crater fill material. MOC image: M0300514. d) Image of ‘oyster shell’ crater in the crater fill material (see also “ring-mold craters” of Kress et al., 2008). MOC image: M0300514. e) Examples of flow features in crater fill within one of the terraces in the crater wall. The white arrows represent the implied flow direction, which is consistent with the local downslope direction. THEMIS image V02146006

Fig. 5. Comparisons of the ejecta deposits of three similarly sized northern mid-latitude craters with that of Sinton crater, showing its unique morphology. a) 90 km Mie crater (at 139°E, 48°N) exhibits ballistic ejecta in the form of chains of secondary craters which extend radially

outward from the ejecta deposit. b) 28 km Arandas crater (-15°E, 42°N) where the ejecta consists of a convex upward annulus of ejecta about one crater radius wide, surrounded by ‘fluidized’ lobate debris sheets. c) 23 km crater at: 15°E, 53°N: exhibits a lobate debris blanket with imprinted radial striations just visible. d) Sinton crater: lacks a distinctive blanket of ejecta material but is surrounded by dense valley networks, which largely emanate from the vicinity of the crater rim. The images are THEMIS infrared composites.

Fig. 6. Dense valley networks and associated fluvial erosional features on the plateau surface to the southeast of Sinton Crater. The boxes correspond to the location of Figs. 7 – 9. North is at the top of the figure. CTX image: P03_002138_2194.

Fig. 7. Valley networks close to the crest of Sinton on the northeastern portion of the plateau. The networks exhibit hanging valleys, small streamlined islands and abandoned valleys. Note that some of the valley forms are poorly defined with several of the networks opening up into broad hollows (light grey). Arrows represent the inferred direction of flow, and the black circles correspond to impact craters. North is at the top of the figure. Subset of CTX image: P03_002138_2194.

Fig. 8. Examples of terraces and raised abandoned valleys within the valley networks along the southeastern portion of the plateau. Arrows represent the inferred direction of flow. North is at the top of the figure. Subset of CTX image: P03_002138_2194.

Fig. 9. Highly complex anastomosing valley network systems on the lower southeastern portions of the plateau. Black circles correspond to impact craters. North is at the top of the figure. Subset of CTX image: P03_002138_2194.

Fig. 10. Relationship between the valley networks and the topographic ridge that extends to southwest of the rim of Sinton crater. The ridge, prominent in the slope map in Fig. 2b, forms a topographic divide, but valley networks are visible on either side of the ridge close to the crest. This configuration, together with the lack of theater-shaped alcoves at the head of the valleys, argues against a groundwater source being responsible for valley formation. The valleys on the western side of the ridge drain into a 5 km wide depression (highlighted in grey), and corresponding valley systems emanate from the down-slope side, arguing for the pooling of water within the depression. 100 m contours derived from MOLA gridded data. North is at the top of the figure. HRSC image 1589.

Fig. 11. Details of the valley networks present along the southern portion of the plateau. The valleys exhibit many of the properties of the networks shown in Fig. 6, including steep flanks, broad floors and anastomosing drainage configurations. They appear to originate in topographic hollows within the outer rim of Sinton crater. As can be seen in Fig. 7, depressions exist across this region of the plateau which are interconnected by the valleys, indicating that the formation of the hollows are contemporaneous with the generation of the valley networks. North is at the top of the figure. HRSC image 1589.

Fig. 12. Interaction between the valley networks and the alcoves along the southern portion of the main plateau. The terrain directly surrounding the alcove is characterized by a smooth surface that is in contrast to the rough-textured unit that comprises the majority of the plateau surface surrounding Sinton crater. The rough texture is most likely the result of ejected material emplaced by the impact that formed Sinton. The valley networks do not open up along the edge of the flanks of the present-day position of the alcove, but instead abruptly terminate at the contact between the two surface types. North is at the top of the figure.

Fig. 13. 500 m wide valley network present on the surface to the west of the main plateau. The valley system is located directly to the west of Sinton crater and constitutes the widest valley network on the plateau. The source of the valley is unclear, as the upstream portions of the valley widen and become indistinguishable from the plateau surface ~1 km from the outer rim of Sinton. Despite the scale of the valley, it exhibits many of the features present in the other networks (Figs 6 and 11), including anastomosing channels, steep walls and a broad valley floor. North is at the top of the figure. THEMIS image V12506004.

Fig. 14. Valley networks and fan-like deposits on the southern interior walls of Sinton crater (a). The sketch map (b) shows that the valleys are discontinuous and that several drain into elongate alcoves toward the bottom of the slopes. At the end of the alcoves, fan-shaped deposits are observed. Evidence for the shallow subsurface flow of water within these deposits is seen in the form of valleys emerging from the downslope margin of several fan deposits (right). In one of these (right), a theater-headed valley begins just downslope of the fan, suggesting that water infiltrated into the fans, causing subsequent shallow groundwater buildup, drainage and downslope emergence, resulting in the formation of the distal alcove and valley. THEMIS image V14016012.

Fig. 15. Conceptual model of the evolution of the valley networks on the surface of the plateau. 1. Anastomosing small-scale valleys form in response to the introduction of surface runoff on the plateau surface. 2. Over time, the runoff becomes concentrated into two of the valleys as the system attempts to maximize efficiency. This causes the valleys to become increasingly incised and widened as a result of the increase in erosion. The other channels are left clogged with sediment and hanging above the main valleys as flow within them ceases. 3. If fluvial erosion continued over longer periods (years – centuries) under relatively constant discharge, the valleys would have become progressively downcut into the plateau and widened in

response to mass wasting along the valley flanks. This would cause erosion and eventual removal of the smaller abandoned valley networks. The preservation of smaller valleys on the plateau surface argues that the valleys never reached stage 3, and thus represent relatively short-lived activity.

Fig. 16. Map of the crater-counting areas on the main plateau. The southern count area represents the portion of the plateau most densely covered in valley networks. Southern surface area is 3201.5 km²; western count surface area is 2030 km². The crater fill area was derived from the mapping of geological units within Sinton crater (Fig. 3); this surface area is 1113 km².

Fig. 17. Crater count plots for (a) the valley networks south of Sinton crater, and (b) the western portion of the plateau within one crater diameter of Sinton crater. The isochrons are plotted according to the Hartmann et al. (2005) convention. (a) Craters were counted on the portion of the plateau that was most densely covered in valley networks, thus the age derived from the crater distributions is considered to be a minimum for the valley forming period. The plots best fit the isochrons close to Hesperian/Amazonian boundary. b) The age correlates with that of a) suggesting that valley networks to the west of Sinton crater also have a minimum age close to the Hesperian/Amazonian boundary. Both plots show that the crater size distributions cross the isochrons at smaller crater diameters (< 500 m) indicating that some process(es) operated to remove the smaller craters.

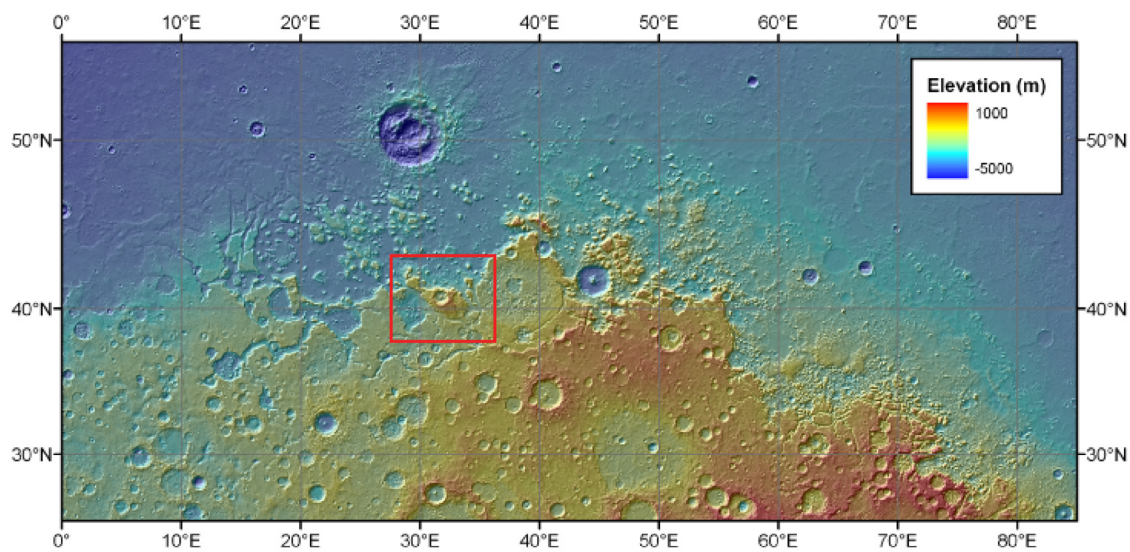
Fig. 18. Crater count plots for (a) the crater fill in Sinton crater, and (b) the surface of main trunk LVF in the south valley (From Morgan et al., 2009). The isochrons are plotted according to the Hartmann et al (2005) convention. Both plots show a similar age of several hundred million years which is consistent with the results of other authors who have dated LVF and LDA deposits in other location across the dichotomy boundary (e.g. Mangold, 2003). The

consistency between the ages of the crater fill deposits with the LVF adds further weight to the interpretation that both deposits are composed of the same material and were deposited during the same event.

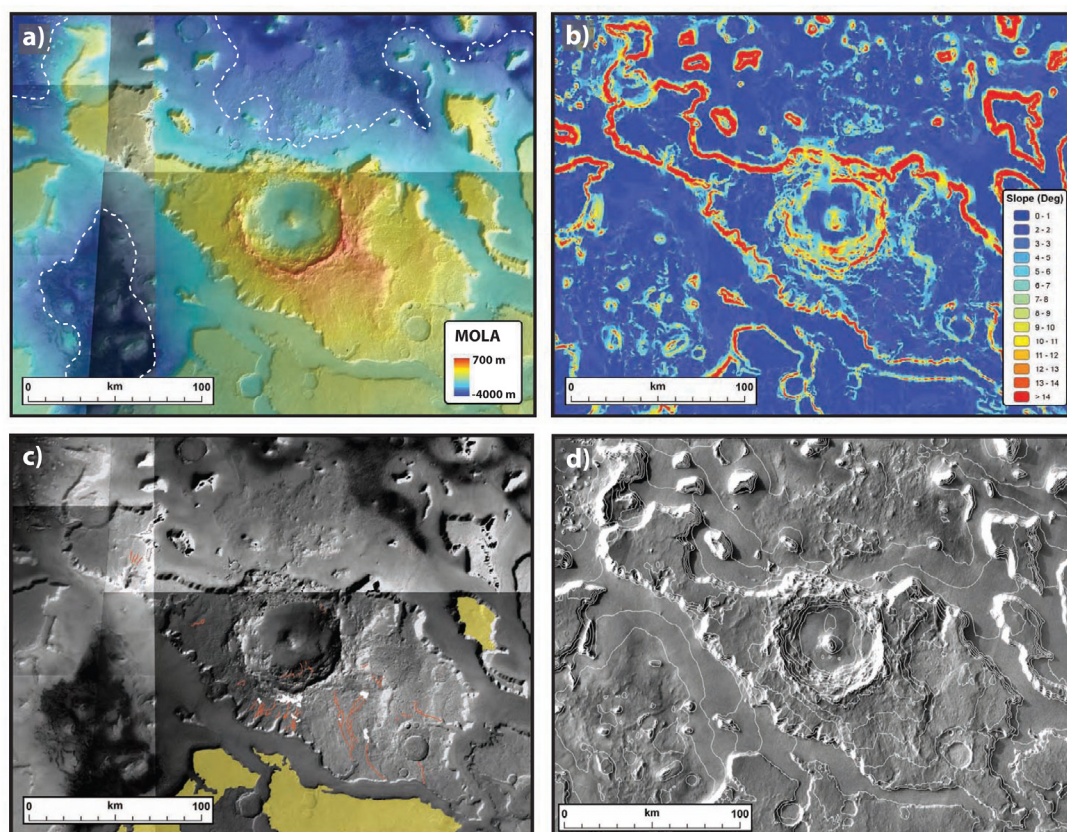
Fig. 19. Model of impact-induced valley formation at Sinton crater depicting a series of idealized west-east cross sections through the main plateau throughout the formation process. Note that the diagrams are vertically exaggerated. a) Ice is deposited during a period of high obliquity, forming ice-rich deposits across the plateau. b) A projectile comes into contact with the surface, causing the excavation of a transient cavity. The ice in contact with the impact along with the projectile and a proportion of the plateau surface are vaporized. c) During the excavation stage ejecta and impact melt leave the crater. The ice deposits are blanketed in hot ejecta which melts the ice, causing the release of water out onto the recently emplaced ejecta. This modifies and erodes the ejecta deposit compared to other relatively similar size craters (Fig. 5). Linear gouges cut into the surface of the plateau by the emplacement of material during the excavation of the crater (secondary crater chains) would have provided natural conduits through which meltwater could flow. This is consistent with channels to the southeast of the crater exhibiting an orientation perpendicular to the crater rim for almost their entire length, apart from short sharp deflections (Figs. 6). d) In the period after the impact the meltwater would have drained off the plateau carving the valley networks and leaving behind the remnants of the ejected material. Some late stage water may have reached the surface through impact-related hydrothermal systems (e.g., Osinski, et al, 2005). Such a process could have maintained activity within some of the valley networks. Modified after Melosh (1989) and Osinski et al, (2005b).

Fig. 20. Geological history of the Sinton crater region interpreted from the results of our analysis placed within the broader history of the dichotomy boundary (as presented by Irwin et al., 2004): 1. Dichotomy boundary formation (Frey et al, 2002). 2. Erosion and burial of the

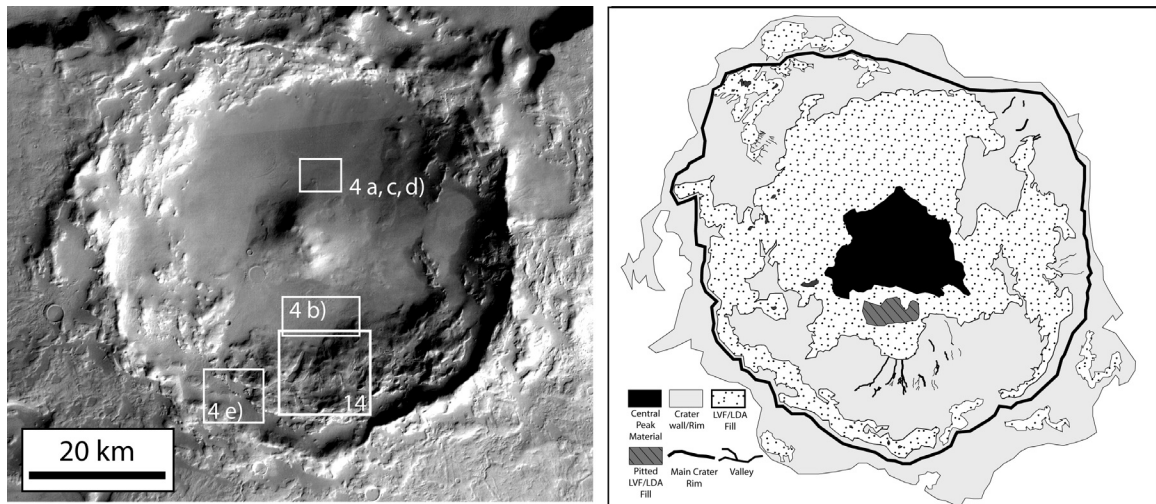
dichotomy boundary (Neukum and Hiller, 1981; McGill, 2000). 3. Formation of large scale valley networks on Mars (dated by Fassett and Head, 2007). 4 and 5. Possible multiple episodes of ice deposition. 6. Pre-impact glacial period. 7. Impact event, forming Sinton crater. 8. Rapid melting of ice deposits and subsequent valley formation. 9. Potential hydrothermal activity around Sinton crater. 10. Possible multiple episodes of ice deposition. 11. Most recent phase of glaciation associated with main trunk of LVF (Morgan et al., 2009). 12. Latest phase of glaciation associated with superimposed lobes (Morgan et al., 2009).



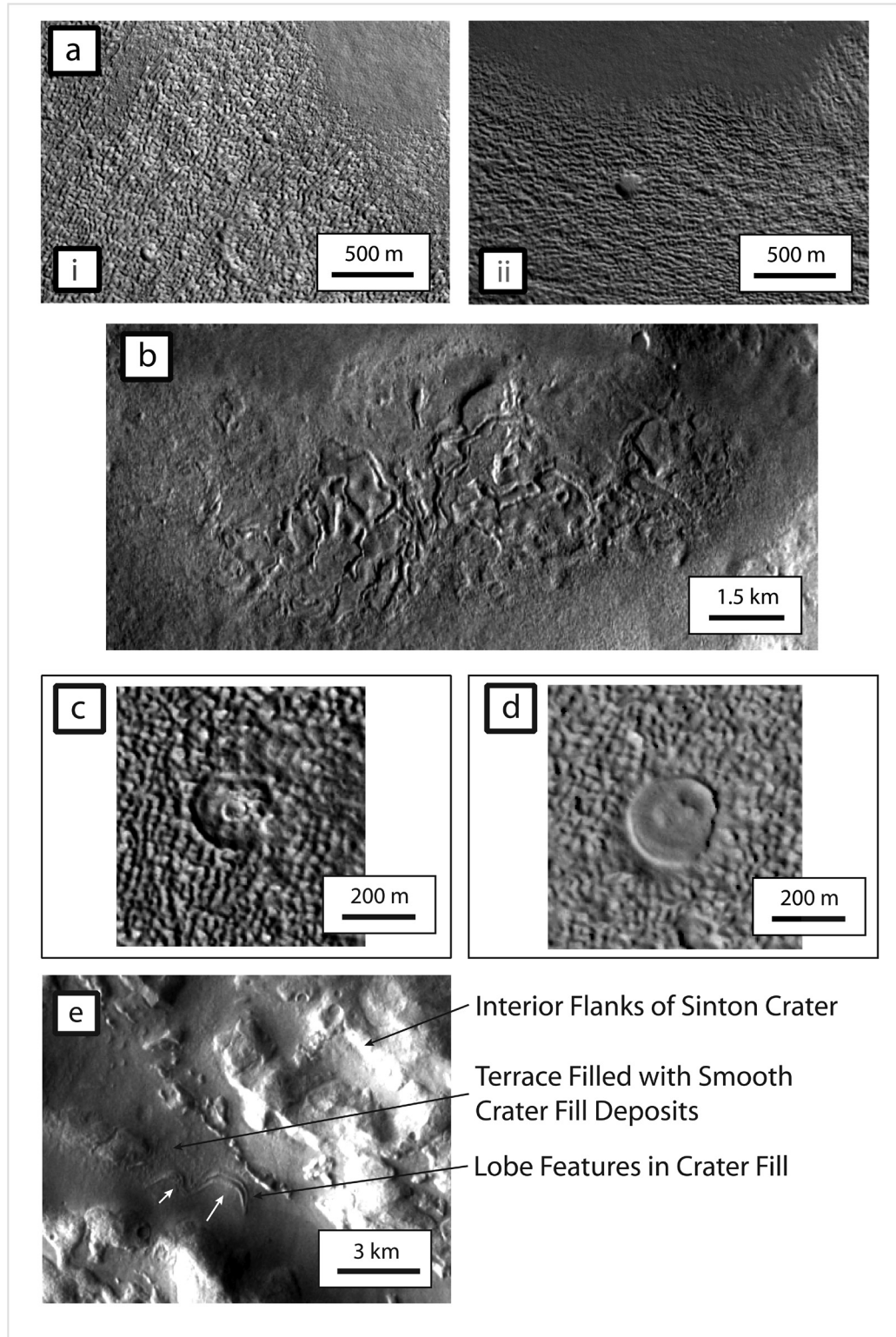
Chapter 1, Figure. 1



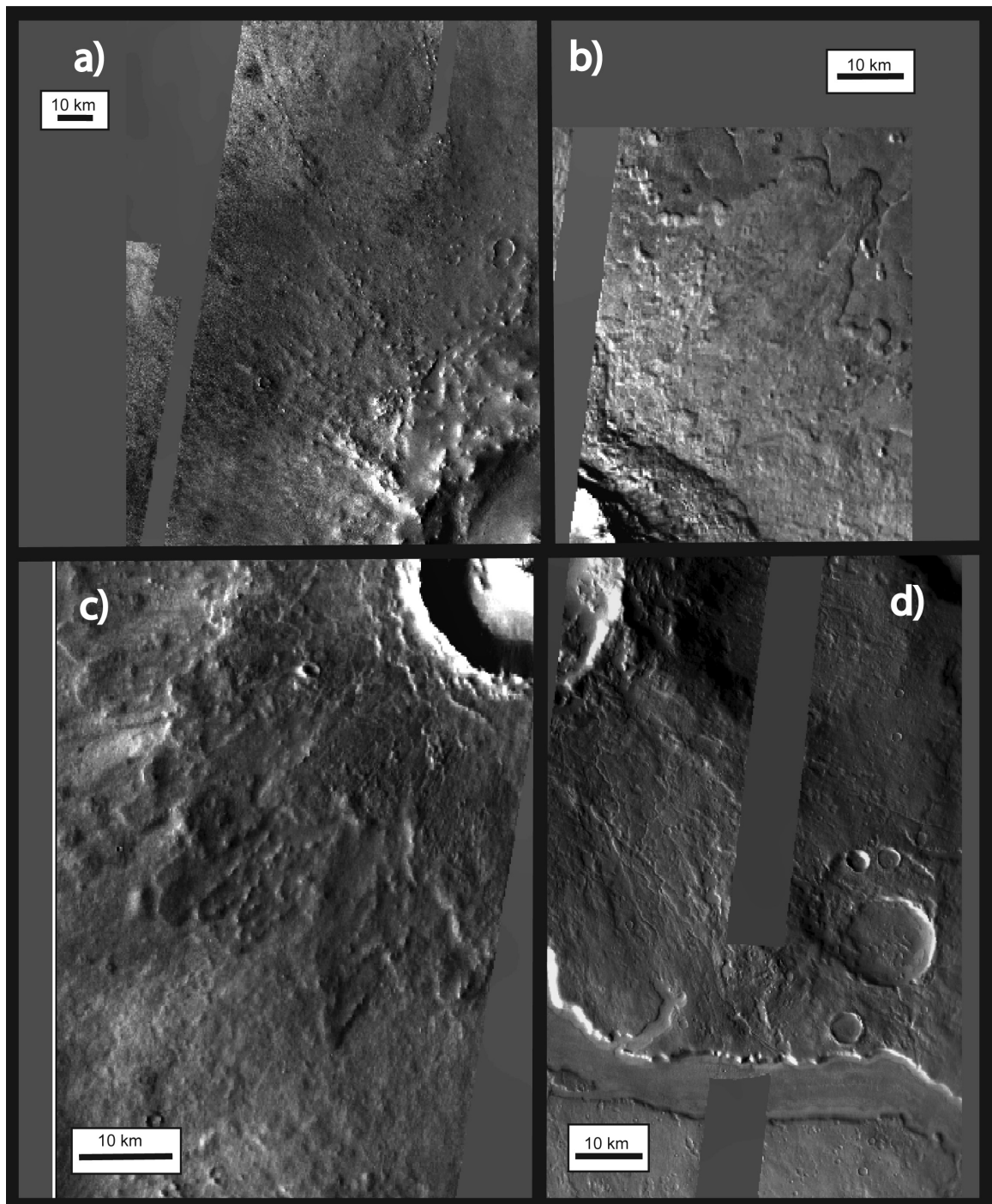
Chapter 1, Figure. 2



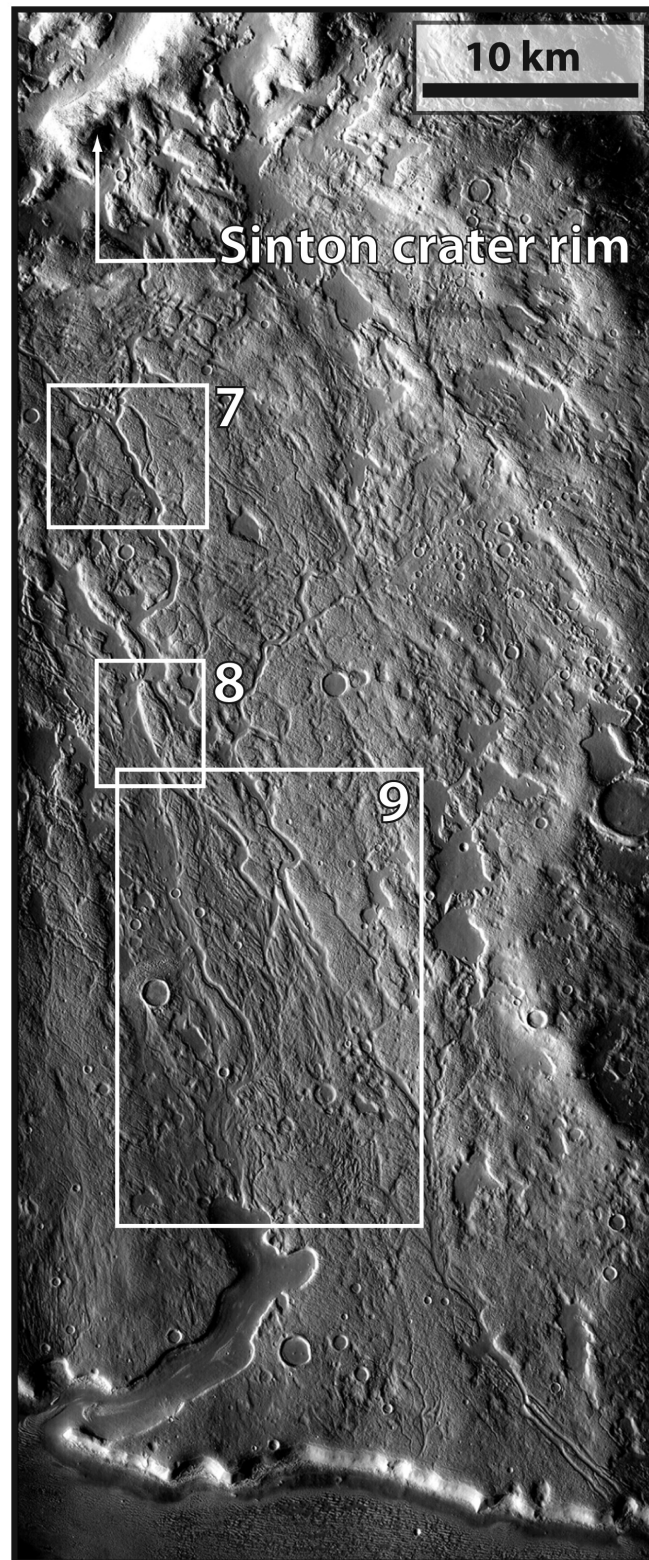
Chapter 1, Figure. 3



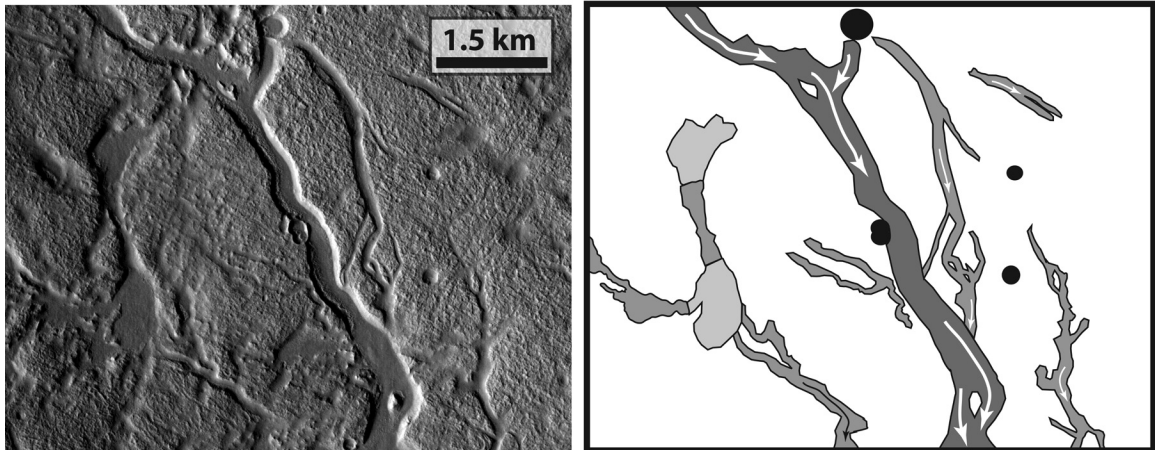
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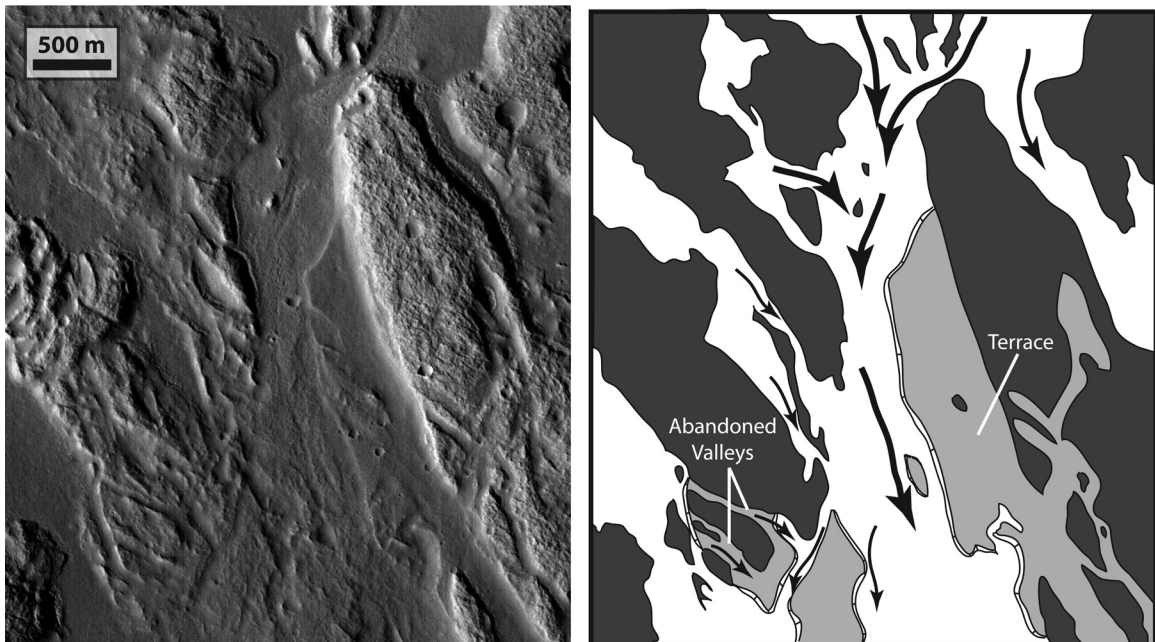
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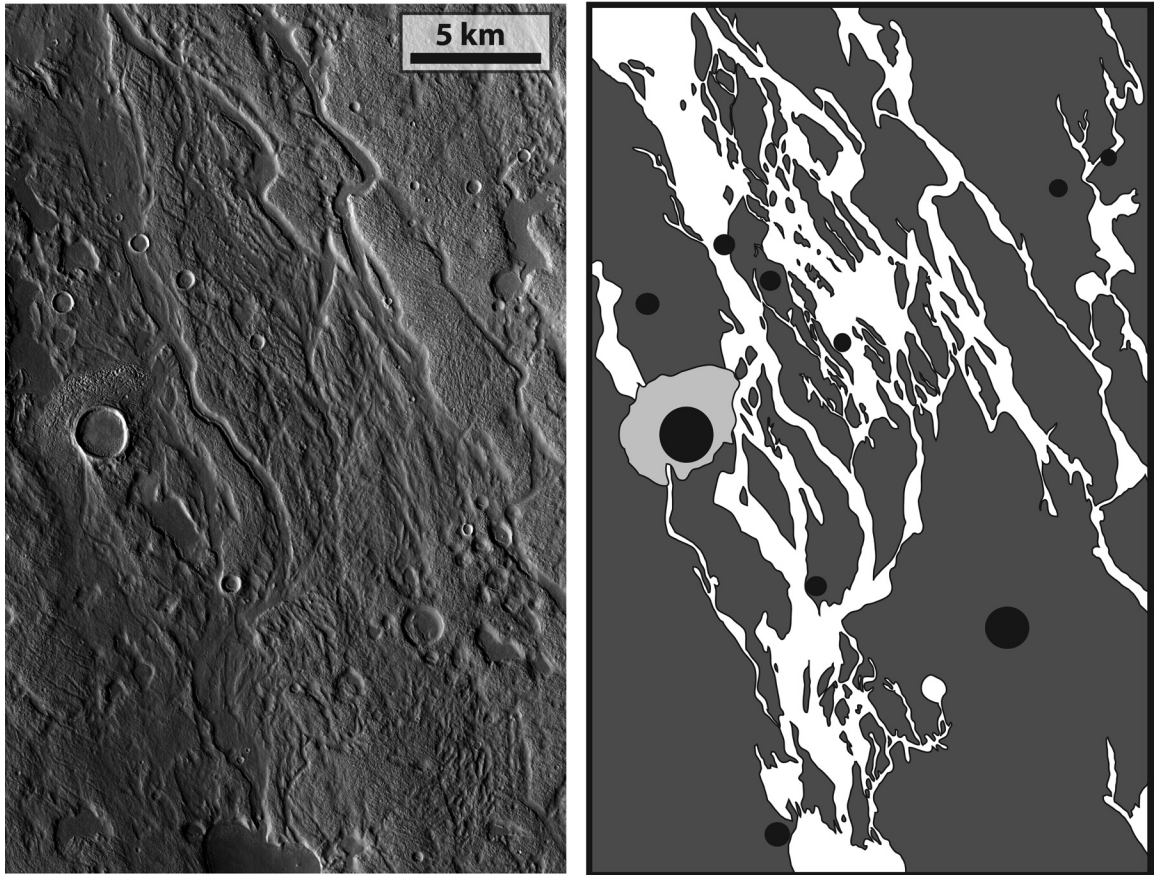
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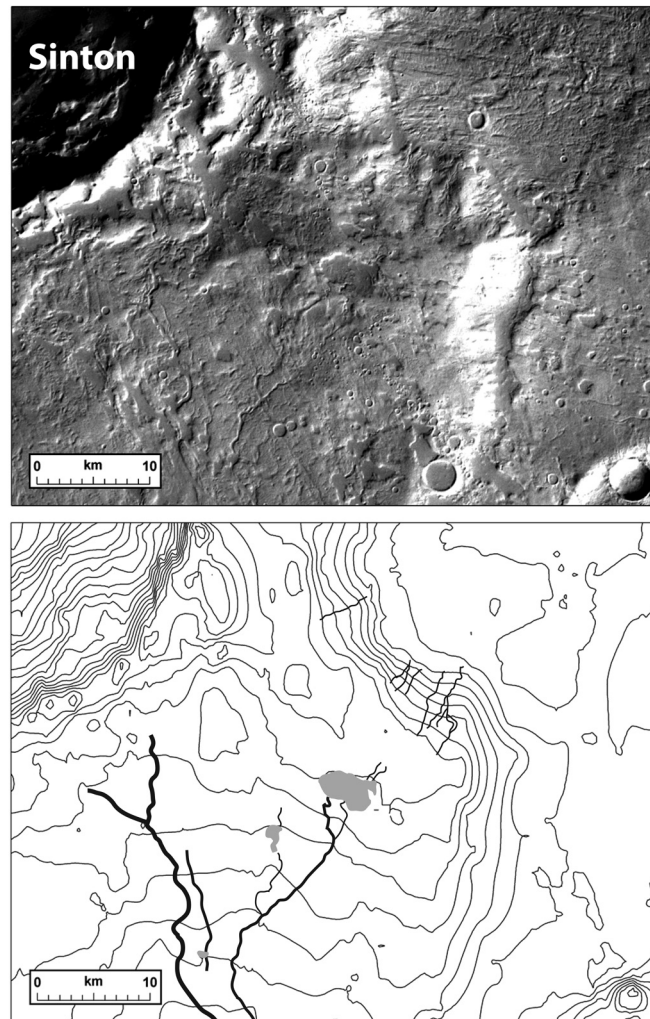
Chapter 1, Figure. 7



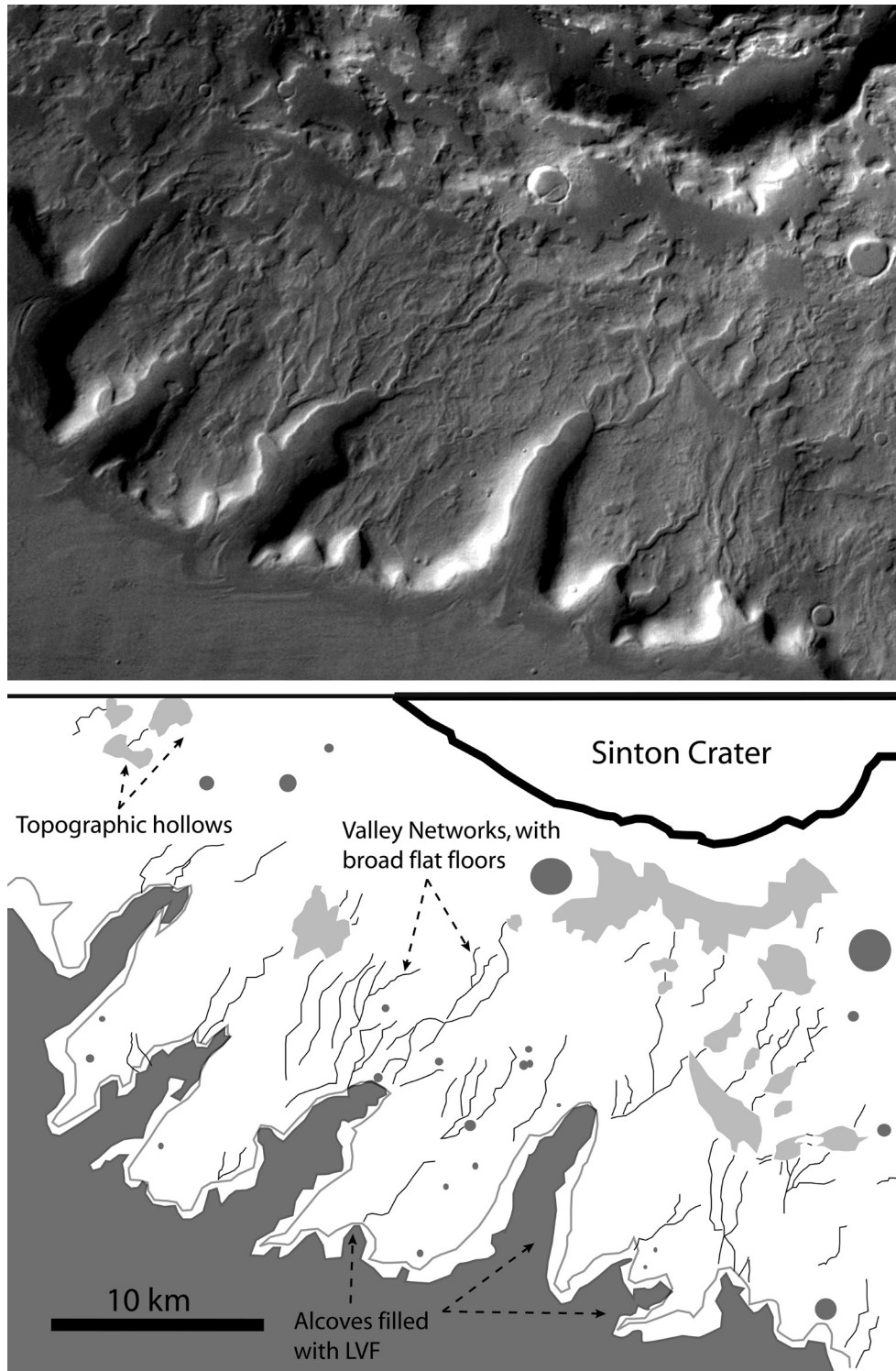
Chapter 1, Figure. 8



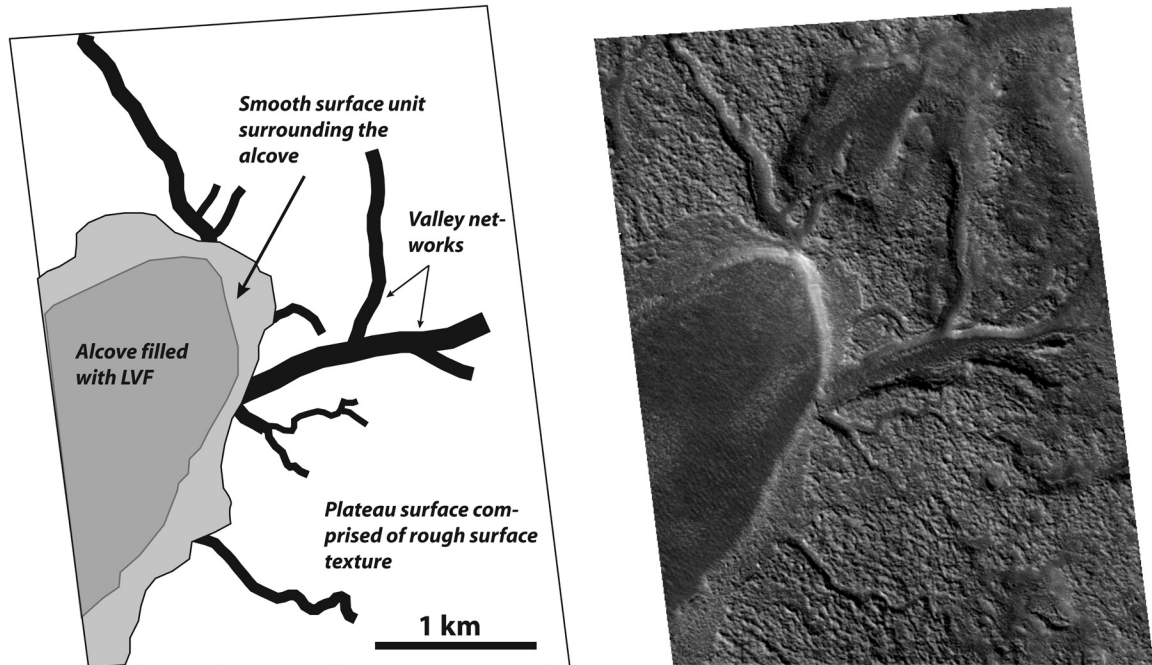
Chapter 1, Figure. 9



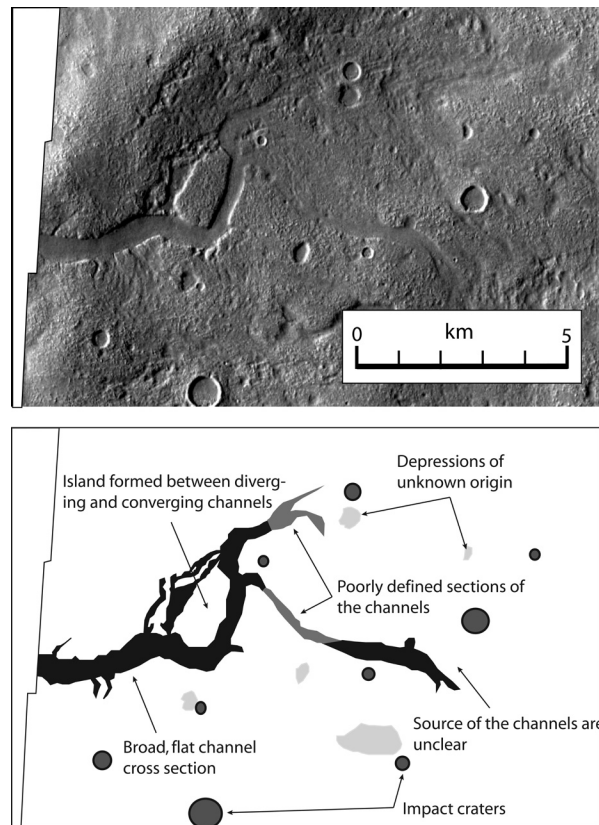
Chapter 1, Figure. 10



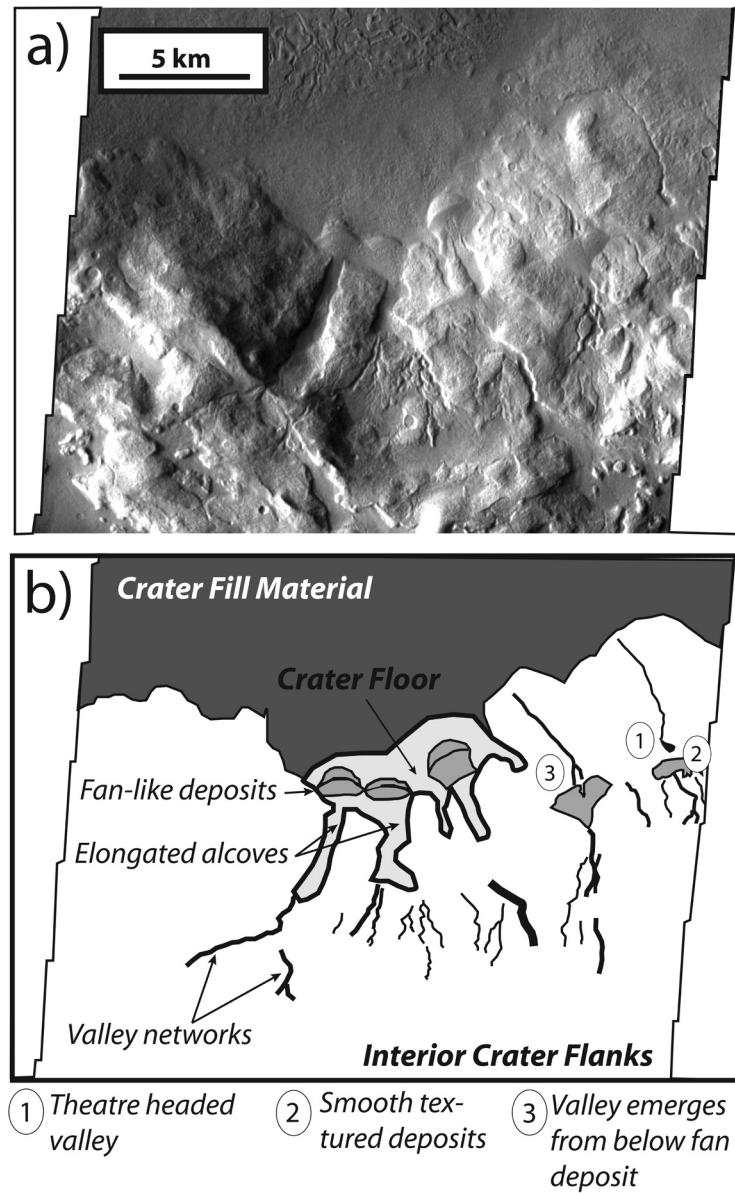
Chapter 1, Figure. 11



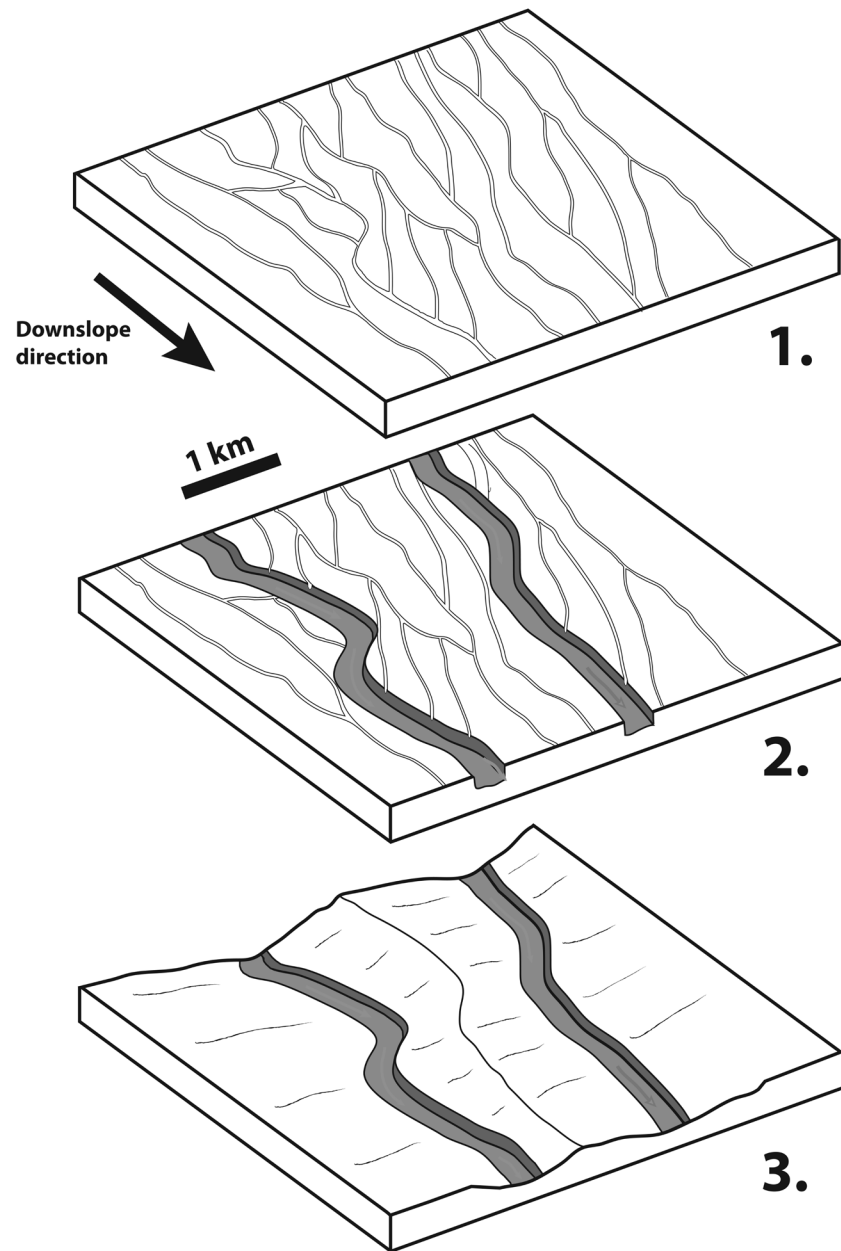
Chapter 1, Figure. 12



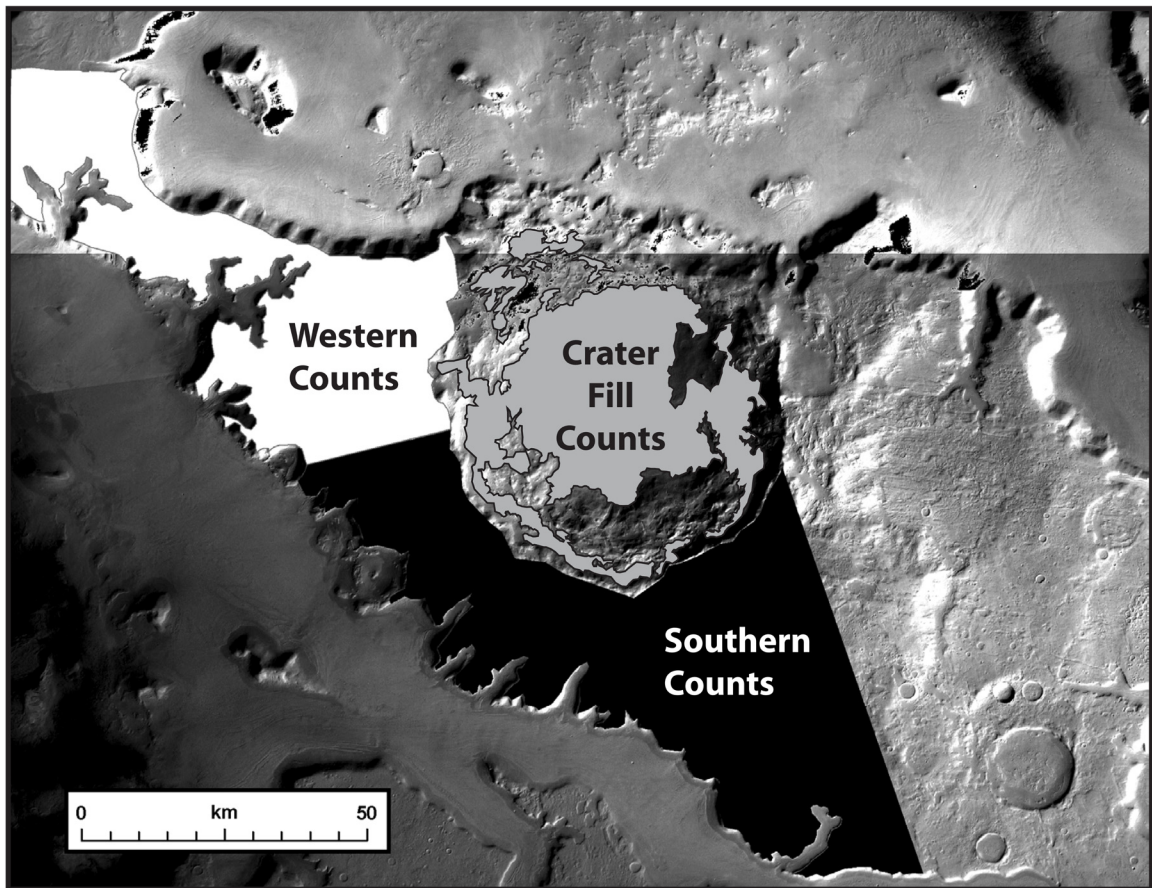
Chapter 1, Figure. 13



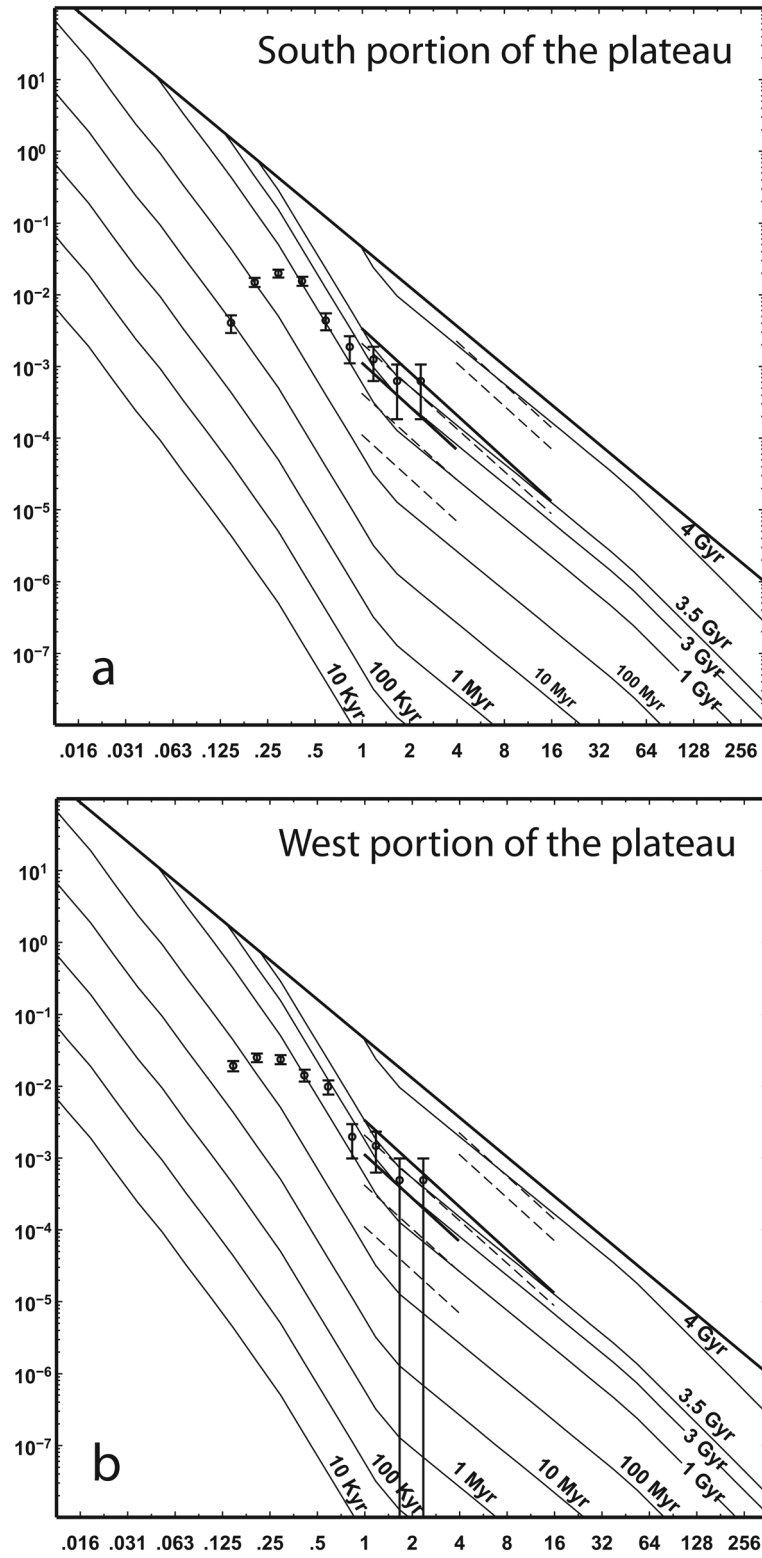
Chapter 1, Figure. 14



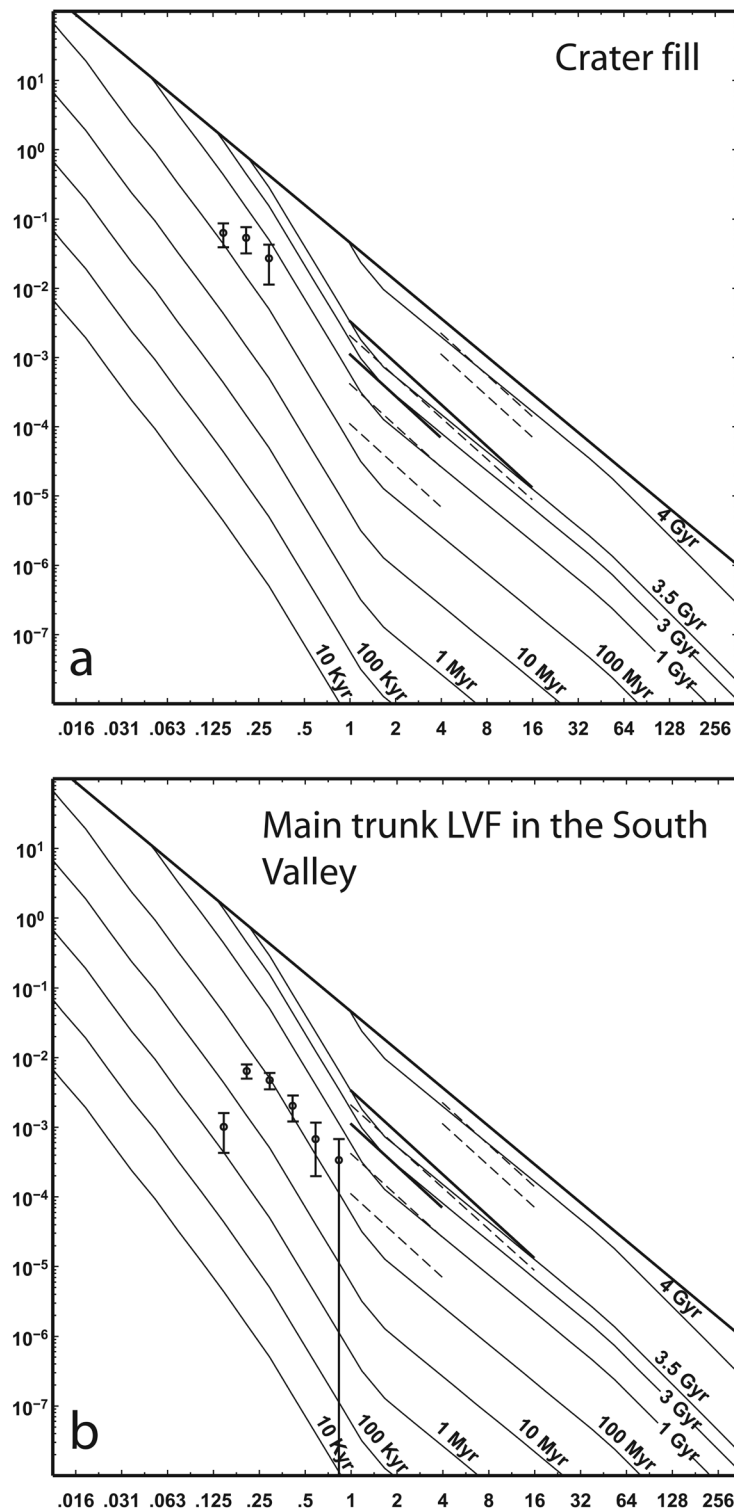
Chapter 1, Figure. 15



Chapter 1, Figure. 16

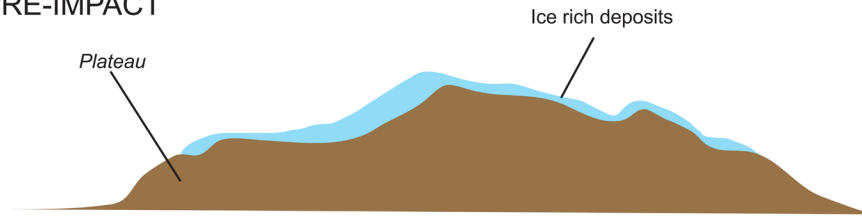


Chapter 1, Figure. 17

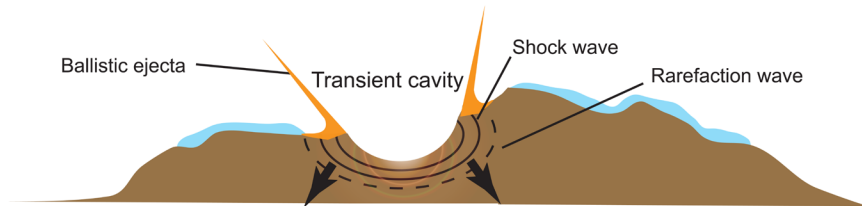


Chapter 1, Figure. 18

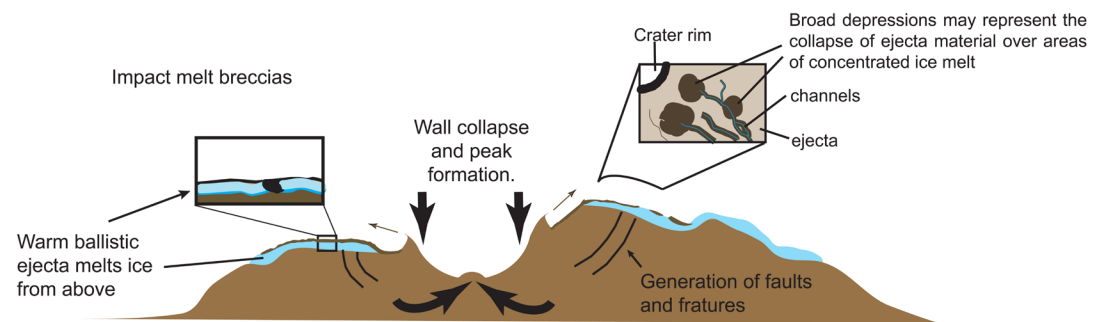
a) PRE-IMPACT



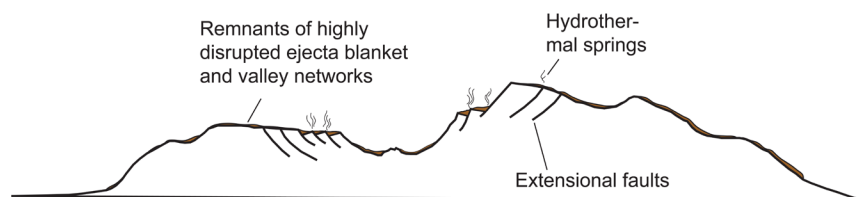
b) CONTACT/COMPRESSION STAGE



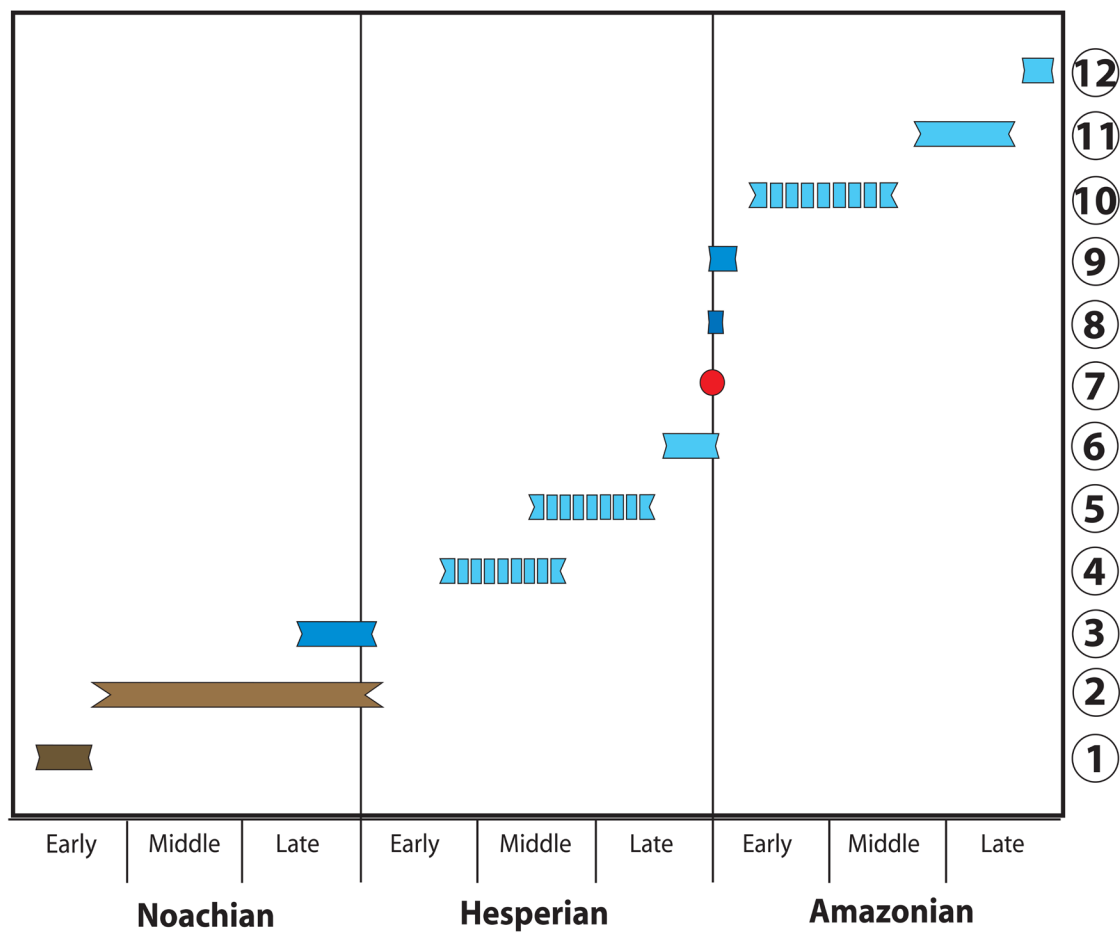
c) EXCAVATION STAGE - INITIATION OF MELTING AND VALLEY FORMATION



d) POST-IMPACT, HYDROTHERMAL ACTIVITY



Chapter 1, Figure. 19



Chapter 1, Figure. 20

Chapter Two

Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA) in the Deuteronilus Mensae Northern Dichotomy Boundary Region, Mars: Constraints on the Extent, Age and Episodicity of Amazonian Glacial Events

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Published in its present form as:

Morgan, G.A., Head, J.W., Marchant, D.R., 2009. Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA) in the Deuteronilus Mensae Northern Dichotomy Boundary Region, Mars: Constraints on the Extent, Age and Episodicity of Amazonian Glacial Events. *Icarus*: doi:10.1016/j.icarus.2008.12.043.

Abstract

In order to assess the nature, degradational processes and history of the dichotomy boundary on Mars, we conducted a detailed morphological analysis of a 70,000 km² region of its northern portion (north-central Deuteronilus Mensae, south of Lyot, in the vicinity of Sinton Crater). This region is characterized by the distinctive sinuous ~2 km-high plateau scarp boundary, outlying massifs to the north, and extensive fretted valleys dissecting the plateau to the south. These features represent the first-order modification and retreat of the dichotomy boundary, and are further modified by processes that form lineated valley fill (LVF) in the fretted valleys, and lobate debris aprons (LDA) along the dichotomy scarp and surrounding the outlying massifs. We use new high-resolution image and topography data to examine the nature and origin of LVF and LDA and to investigate the climatic and accompanying degradational history of the escarpment. On the basis of our analysis, we conclude that: 1) LVF and LDA deposits within the study region are comprised of the same material, show integrated flow patterns, and originate as debris-covered valley glaciers; a significant amount of ice (hundreds of meters) is likely to remain today beneath a thin cover of sublimation till. 2) There is depositional evidence to suggest glacial highstands at least 800 m above the present level, implying previous conditions in which the distribution of ice was much more widespread; this is supported by similar deposits within many other areas across the dichotomy boundary. 3) The timing of the most recent large-scale activity of the LDA/LVF in this area is about 100 - 500 million years ago, similar to ages reported elsewhere along the dichotomy boundary. 4) There is evidence for a secondary, but significantly limited phase of glaciation; the deposits of which are limited to the vicinity of the alcoves; similar later phases have also been reported elsewhere along the dichotomy boundary. 5) Modification of the fretted valleys of the dichotomy boundary has been substantial locally, but we find no evidence that the Amazonian glacial epochs caused retreat of the dichotomy boundary of the scale of tens to hundreds of kilometers. Our findings support the results of an analysis just to the east of

the study region and of studies carried out elsewhere along the dichotomy boundary that find further evidence for the remnants of debris-covered glaciers and extensive valley glacial land systems.

1. Introduction

The Martian dichotomy boundary divides the lowland northern plains from the heavily cratered southern highlands, and is marked by a near-global escarpment that exhibits several kilometers of relief (Fig. 1). The origin of the northern plains and the formation and evolution of this escarpment remain among the most enigmatic chapters in the geological history of the planet (Watters and McGovern, 2006). The highlands flanking the escarpment are considered to be typical of the material that comprise Noachis Terra to the south and are interpreted to be composed of both sediments and volcanic material that has been mixed and reworked by impacts (Tanaka et al., 2005). The northern lowlands have served as a repository for both sediments shed from the surrounding ancient highlands and for volcanic flows and deposits from sources within and near the lowlands (Tanaka and Scott, 1987; Tanaka et al., 2005). The derivation of hypotheses and models to explain the boundary relies on successfully constraining the original position and scale of the escarpment at the time of its formation. This can only be achieved with a comprehensive understanding of the degradation processes operating along the boundary and by making assessments of the level of modification it has undergone, permitting the lateral extent of retreat over geologic time to be mapped and identified (Watters and McGovern, 2006).

The most topographically distinctive region of the dichotomy boundary is located along its northernmost margins (30-50°N) at Deuteronilus-Protonilus Mensae (Fig. 1). This area is dominated by fretted terrain that is defined by sinuous and linear valleys that extend from the southern highlands and widen into the northern plains, isolating plateaus and mesas that become progressively smaller to the north (Sharp, 1973). The modification of this portion of the boundary

has in part been attributed to two distinct degradational features; Lobate Debris Aprons (LDA) that originate from and surround isolated mesas and the escarpment flanks, and Lineated Valley Fill (LVF) which is found to line the floors of the fretted valleys (Squyres, 1978). Three general hypotheses have been proposed for the origin of LVF and LDA: 1) Mobilization of ancient ice-rich highland regolith, in part helping to create the fretted valleys, and then finally leading to the formation of LDA-like features (Lucchitta, 1984; Carr, 1996); 2) Vapor diffusion of atmospheric water vapor into talus, providing a lubrication mechanism, and permitting viscous flow along flanks of valley walls and massif margins, has also been proposed, although the exact nature of the amount of ice has been highly debated (e.g. Carr, 1996, Mangold, 2003); 3) Following earlier suggestions (Lucchitta, 1984), more recent studies have hypothesized that LDA and LVF are largely glacial in origin (Head and Marchant, 2006a, 2006b; Head et al., 2006a, 2006b; Kress et al., 2006), a hypothesis supported by observations from research on debris-covered glaciers in terrestrial analog sites, especially the Antarctic Dry Valleys (Marchant and Head, 2004; 2007). In this scenario, LVF and LDA represent debris-covered glaciers that formed in an earlier period of the history of Mars when climate change caused snow and ice to accumulate preferentially in these regions. According to this model, regional snow and ice accumulation caused glacial flow in valleys (LVF) and at the margins of massifs (LDA), and debris from adjacent massifs and valley walls created a cover of sublimation till that protected the buried ice from sublimation when climate conditions changed. In contrast to the ancient ground ice and viscous flow hypotheses, this scenario predicts significant amounts of primary ice in LVF/LDA formation, with much of it remaining today (Marchant and Head, 2007; Dickson et al., 2008).

Despite much recent progress, several uncertainties remain in the understanding of the degradational history of the dichotomy boundary and the processes that form the LVF and LDA:

- 1) What is the proportion of ice and debris involved in the initial stages of formation of LDA and LVF and how much remains today?
- 2) What were the lateral and aerial extent of the conditions that produced the LDA-LVF at the dichotomy boundary?
- 3) What was the timing and duration of

these periods of accumulation and flow and how do these help distinguish models of formation?,

4) What were the mechanisms for accumulation of ice and snow that might have caused these features and the general link to martian paleoclimate conditions?, and 5) Is there evidence for multiple periods when LVF and LDA may have occurred in these regions?

In this study we concentrated our efforts on the fretted valleys surrounding the 63 km diameter Sinton impact crater formed on a plateau on the northern edge of the dichotomy boundary at 32°E, 39°N (Fig. 2). This region was chosen because it represents a high-latitude portion of the dichotomy boundary, it includes examples of both LVF and LDA, it contains several different orientations of fretted valleys, it has abundant LDA-containing massifs, and it is adjacent to a 30,000 km² region containing a regionally integrated system of LVF (Head et al., 2006b). Through assessment of an array of spacecraft data including MOLA altimetry, HRSC DTMs, and THEMIS, HRSC, MOC, CTX and HiRISE high-resolution images, we analyzed LVF and LDA within the region in detail to address the questions outlined above.

2. The Evolution of Theories of LDA/LVF Generation and Critical Observations

Early hypotheses pertaining to the processes responsible for the LDA called upon the actions of frost creep or gelifluction, operating on ancient ice-rich, permafrost-like regolith during earlier periods of martian history when increased geothermal gradients caused mobilization and flow (Carr and Schaber, 1977). Observations supporting this hypothesis would be: 1) a relatively ancient age of LDA and LVF, 2) distinct continuity between plateaus undergoing mobilization and the resulting lobes 3) evidence for associated melting and water flow, and 4) modest amounts of ice involved (pore ice and secondary ice). As new data became available regarding the thickness of these features (0.5 – >1 km), their apparent youthfulness, and the inability of the current/recent climate to permit thawing to significant depths, frost creep was considered an unlikely candidate for such large landforms (Squyres, 1978).

Squyres (1978) proposed that the LDA were comparable to terrestrial rock glaciers, in that they were mass wasting products (talus piles) lubricated by the accumulation of ice in pore spaces permitting them to flow. Snow (and snowmelt) is typically the source of ice for such landforms on Earth, but diffusion of atmospheric water vapor was considered to be a more likely source for the ice in the martian case (Squyres, 1978). Considering the morphological similarity between LDA and lineated valley fill (LVF), LVF was assumed to be the result of two or more LDA flowing from opposing valley walls, and converging in the valley center, completely covering the valley floors (Squyres, 1978). LVF is characterized by along-valley lineations composed of ridges and grooves that had been identified in the original Viking images and were subsequently explained by Squyres (1978) to be the manifestation of compressional forces as the LDA converged within the middle of the valley. Observations supporting this hypothesis would be: 1) localized patterns of LDA convergence to produce LVF, 2) minimum down-valley flow, 3) morphometric parameters indicative of pore ice and small amounts of secondary ice, rather than large ice volumes, (e.g. cross sectional profiles indicative of viscous flow associated with high debris content rather than ice flow) and 4) irregular down-valley topographic gradients in fretted valleys, indicative of localized accumulations from adjacent valley walls (Carr, 2001).

In contrast to these interpretations, Lucchitta (1984) initially noted a down-valley flow component evident in some examples of LVF, and compared these patterns to those seen in Antarctic glaciers. On the basis of increased knowledge of Antarctic analogs (see Marchant and Head, 2007), this hypothesis has recently been further developed by Head et al. 2006a,b) who argue that LDA represent debris-covered glaciers and that LVF represents the remnants of debris covered valley glacial landsystems (Marchant and Head, 2007). In this scenario, regional ice and snow accumulation results in plateau icefields and valley glaciers that converge and flow downslope in fretted valleys to create regional glacial landsystems; the current deposits represent the remnants of previously much more extensive glacial landsystems (Dickson et al., 2008), with remnant ice currently protected by a cover of sublimation till derived from adjacent scarps. LDA

represent localized debris-covered glacial remnants surrounding isolated massifs. GCMs developed for Mars have demonstrated that substantial snow and ice can be deposited along the dichotomy due to the redistribution of volatiles during high obliquities (Madeline et al, 2007). Simulations of the orbital history of the solar system demonstrate that Mars has undergone significant variations in obliquity throughout its history which has included excursions greater than 80° (Laskar 2004). Observations supporting the glacial hypothesis would be: 1) cirque and valley glaciers in the proximal areas of LVF, 2) convergence and flow to create regional integrated patterns of LVF, 3) morphometric parameters indicative of large ice volumes, 4) down-valley topographic gradients in fretted valleys indicative of regional, not local, flow.

3. Methodology and Approaches to Studying LVF/LDA

3.1 The Study Region

The study region is centered at 30°E , 40°N , within Deuteronilus-Protonilus Mensae. This location was selected because it was situated between two previously studied regions, one to the west (Head et al., 2006a) and one to the east (Head et al., 2006b), and thus enabled us to extend the analysis of the most prominent portion of the dichotomy boundary. The LVF system within the study area extends throughout a region of $70,000 \text{ km}^2$, and is best expressed within a $\sim 15 \text{ km}$ wide valley formed between the main plateau on which Sinton crater is situated and the cratered uplands (Fig. 2). LVF is also present within connecting smaller tributary fretted valleys and breached craters that are located to the south of the main plateau (Fig. 2). Large areas of LDA are present within the study region, predominantly along the northern flanks of the main plateau and around the walls of the 130 km basin located in the western part of the study region. Figure 3 illustrates the nature of the morphology of LDA, which consist of low slope angles ($\sim 1^\circ$) along the main body of the aprons, steepening to $< 5^\circ$ towards the terminus (identifiable by the yellow and red shades and highlighted by the arrows in Fig. 3), forming a convex cross-sectional profile

(see Fig. 4). Such a profile is consistent with LDA located across the northern dichotomy boundary region (Squyres, 1978; Li et al, 2005).

3.2 Data Sets and Analysis

The investigation was conducted through the compilation of a Geographical Information System (GIS) database comprised of the relevant martian data sets for the study region. The database consisted of: (i) topographic data from both 128 pixel/degree gridded Mars Orbiter Laser Altimeter (MOLA) data and High Resolution Stereo Camera (HRSC) stereo data (200 m/pixel); (ii) visible image datasets including: (1) (HRSC) images (18 m/pixel), 2) Thermal Emission Imaging System (THEMIS) visible images (19m/pixel), 3) Mars Orbital Camera (MOC) narrow angle images (1.5 – 12 m/pixel), 4) Context Imager (CTX) images (8 m/pixel) and 5) HiRISE images (0.3 m/pixel); and (iii) thermal infrared data from daytime and nighttime THEMIS infrared images (band 9, 100 m/pixel). ESRI's ArcMap (9.2) provided the GIS platform, which in addition to dataset management was also utilized to produce slope maps and topographic profiles from the interpolated MOLA data and conduct crater counts, all of which contributed to the research process.

4. Observations and Interpretations of LDA/LVF in the Study Region

4.1 Alcoves and Tributary Valleys and the Source of LVF/LDA

We found that alcoves and tributary valleys along the fretted valley margins and the scarp margin were a prevalent feature throughout the study region (Figs. 1-6). On the basis of our mapping of regional LDA/LVF patterns (Fig. 5), it is clear that these tributary valley/alcoves serve as a source region for individual LVF lobes. Within the study region the alcoves are found not only in conjunction with fully integrated LVF within the main valley that runs east to west to

the south of the main plateau, but also along plateau flanks that are lined with LDA (Fig. 2). The occurrence of lobate material sourced from within alcoves that coalesce and integrate with both LDA and LVF systems argues that both landform types (LDA and LVF) originated from the same process (Head and Marchant, 2006b).

The majority of alcoves were found along the southern flanks of the main plateau. They exhibit a range of scales from small enclaves (several kilometers across) to complex systems consisting of several tributary alcoves (Fig. 7), which together formed tributary valleys to the larger trunk valley system (Fig. 7). Incision within the plateau by these larger alcoves can be significant, of the order of tens of kilometers and to the extent that the plateau is almost completely breached along its thinnest sections (Fig. 7). This may provide insight into the means by which mesas form along the dichotomy boundary, as plateaux such as the main one in the study region undergo increasing fragmentation or ‘mesa-ization’ (Head et al, 2006b). There is a wide range in scale and complexity of alcoves and their distribution produces a fractal pattern within the plateau flanks. This is due to smaller, lower order alcoves occupying the space along the plateau walls between the larger alcoves and also within the alcoves themselves, forming tributaries and small enclaves. Differences in the development of alcove forms may therefore reflect differences in maturity, lithology, or be due to the pirating of preexisting structures within the plateau and surrounding mesas. Tectonic (McGill and Dimitriou, 1990) and fluvial activity (Carr, 2001) during the late Noachian and early Hesperian are thought to have generated the initial alcove landforms, which may have been capable of providing the necessary shelter for the accumulation of ice during the later periods of glacial activity associated with the most recent Amazonian ice ages (e.g. Head et al., 2006b). Therefore, it may be difficult to ascertain the exact contribution that processes related to glacial activity alone have provided to the erosion of the dichotomy boundary. Nevertheless, the very large number of modified alcoves and enclaves which occupy almost every kilometer of the plateau flanks, and those of the surrounding mesas, is testimony to the level of local erosion that the dichotomy boundary has experienced due to

apparent glacial-associated processes.

Longitudinal lineations consisting of orientated ridges run the length of the majority of individual LVF lobes and are aligned with the prominent direction of flow (Figs. 7 - 10). At the mouth of the alcove in Fig 8, the lineations are observed to form chevron patterns in the downslope directions. On terrestrial glaciers, such 'zigzag' patterns are observed as contorted medial moraines and supraglacial debris resulting from folds developing in the ice due to compressional and shear forces. In the case of Fig 8, such folding can be explained by comparable compressional stresses acting on the body of the LVF. In the complex alcoves, materials within the individual enclaves merge with the LVF within the main trunk to form lobes that become wider as a function of distance downslope. A particularly good example of the interaction of such material is seen in Fig. 9, where two separate lobes develop and converge together at the opening of the alcove system. This is made particularly apparent by the individual longitudinal lineations that run along the central portions of LVF within the enclaves. These merge together at the mouth of the alcove system and form a single series of parallel ridges (Fig. 9) that are similar to the broad ridges present on the surface of LVF lobes emanating from simple alcove systems (Fig. 10).

The point of contact between lobes emanating from the alcoves and the major LVF and LDA systems is one of convergence, as one deposit merges into the other, further suggesting that they all consist of the same material, and reflect integrated flow (Fig. 7). In some circumstances concentric ridges perpendicular to the direction of flow can be observed in the LDA downslope of the emerging lobe (Fig. 9). As the lobe material becomes increasingly integrated within the LDA in the downslope direction, the ridges become progressively less defined to the point that they can no longer be identified. This occurs at a distance of ~10 km from the alcove mouth. Such morphology is also apparent with integrated debris covered glacial systems on Earth, such as Beacon Valley in the McMurdo Dry Valleys of Antarctic (Fig 9), an important analog for geological processes on Mars (Marchant and Head, 2007; Levy et al., 2006; Shean et al., 2007).

For example, a similar relationship is seen where the Mullins Valley debris-covered glacier integrates with debris-covered ice within central Beacon Valley (Fig. 9). Concentric ridges resulting from compression associated with the advance of Mullins are highlighted by the deposition of windblown snow within the troughs between the undulations that occur down-slope of the entrance of Mullins valley.

These relationships strongly suggest that LVF and LDA systems originated from the same source regions and environments (alcoves and tributary valleys), and grew over time until enough material had emerged from the alcoves to generate the LVF that fills the fretted valleys. Such observations and interpretations are consistent with the investigations carried out to the east by Head et al. (2006b) and other localities across the dichotomy boundary, including Protonilus Mensae (Levy et al, 2007), Mamers Valles (Kress et al., 2008) and Coloe Fossae (Dickson, 2008). Together, these observations are consistent with the debris-covered glacial model of LVF/LDA emplacement. The recent acquisition of SHARAD (SHallow RADar) subsurface radar data in an extensive area of LVF and LDA (Head et al., 2006a, b) strongly suggests that almost pure water ice underlies a thin debris-rich surface (Plaut et al., 2008). This adds further support to the interpretation that the two-landform types are the product of debris-covered glacial activity.

4.2 Evidence for Integrated Systems of LVF/LDA

From observing the apparent flow patterns in HRSC, THEMIS and MOC high-resolution images throughout the entire study region (Fig. 5), it has been possible to map integrated systems of LVF and LDA across this area of the dichotomy boundary. Such mapping techniques have already been applied successfully to the LVF systems to the east of the study region (Head et al, 2006b). This mapping revealed an integrated flow system extending in excess of 200 km in length, covering an area of $\sim 30,000 \text{ km}^2$. In our study area, integrated LVF systems are especially

apparent within the LVF and LDA, surrounding the western extent of the main plateau (Fig. 5).

The convergence of surface lineations reveal the constriction of the flow of LVF. A prominent example of this is seen in Figure 6a where the downslope flow of LDA has been constricted through the opening of a topographic ridge to produce a bulbous lobe of LDA ~5 km wide on the downslope side. Flow constrictions are also apparent between broader-scale obstacles, as can be seen by the relatively more subtle convergence surface features between the mesa and the western edge of the main plateau (Fig. 6b). In contrast to this, flow also diverges around obstacles such as mesas (Fig. 5).

The integrated nature of the LVF/LDA systems within the study area is morphologically similar to terrestrial networks of connected valley glaciers that form almost web-like patterns within regions of high relief, that are too dissected to support an ice cap (e.g. the current valley glacial landsystems of the McMurdo Dry Valleys, Antarctica, Marchant and Head, 2006). These terrestrial systems have been called *transection glaciers* by Benn and Evans (1998), and can grow sufficiently large that they can overcome local drainage configurations. Evidence of flow is observed from opposing ends of valleys that have high stands at points midway along the center of the valley. This is the case in the large valley to the south of the main plateau in the study area, in which the highest elevation of its floor is in the center, as opposed to one of its ends (Fig. 2). The material flowing from the alcoves into this valley curve away from the point of maximum elevation (Fig. 5), supporting the interpretation that along-valley flow is occurring at either end of the valley. Squyres (1978) cited the prevalence of lineated valleys across the entire dichotomy boundary area as evidence that LVF formed from simple convergence of flow from both sides of the valley, with minimal to no lateral flow. Identification of the alcoves as sources of LVF, however, suggests that glacial processes can indeed explain such situations.

MOLA-derived, along-slope profiles of the LVF and LDA systems (Fig. 4) reveal that all of the deposits correspond to similar lobate cross-sections which are characteristic of LVF/LDA bodies throughout the region (Mangold and Allemand, 2001; Turtle et al., 2003; Li et al., 2005).

The LVF slope within the valley to the south of the main plateau tilts towards the west at an angle of $\sim 1^\circ$, but steepens rapidly to $>4^\circ$ where the LVF terminates into a large basin (Fig. 2), producing a convex-up profile with a relatively steep front that is reminiscent of the fronts of many terrestrial glaciers (Figs. 3, 4). Similar profiles are also observed in the LDAs that emanate away from the northwestern portion of plateau and into the northern plains (Fig. 4). Investigations of the profiles of other LDAs across the dichotomy boundary compare well to plastic models of deformation indicative of the presence of high concentrations of ice (Mangold and Allemand, 2001; Turtle et al., 2003; Li et al., 2005). The north-south trending profiles, which cross LVF/LDA along the western portion of the study region, also reveal that the small mesas to the north have similar surface elevations to each other and to the main plateau. Such topographic traits were observed in the area immediately to the east (Head et al., 2006b). Such profiles are consistent with glacial erosion alongside mesas, reducing their overall volume but leaving their summit elevations intact; over time this has resulted in the gradual ‘mesa-ization’ (Head et al., 2006b) of the northern margin of the dichotomy boundary.

4.3 Evidence for Post-Flow Modification of the LVF/LDA Systems

Most workers agree that the fundamental flow-like patterns observed in the LDA/LVF at Viking, THEMIS VIS and HRSC resolution reflect primary surface textures that date from the emplacement of the deposits and are caused by their lateral motion and flow (e.g., Mangold, 2003; Head et al., 2006a, b). Studies conducted using the highest resolution data sets (MOC and more recently CTX and HiRISE) however, have revealed the LVF/LDA surface environment to be comprised of a collection of complex textures, that have in part been formed by more recent modification (e.g. Mangold, 2003; Levy et al, 2009). In a review of observations made by the Mars Global Surveyor spacecraft of the dichotomy boundary, Carr (2001), utilizing high resolution MOC images was able to further characterize the fretted terrain environment and found

it to consist of three main components: 1) a steep upper slope, where bedrock may be visible, 2) intermediate units (IU) which appear smooth, and may have faint lineations in the downslope direction, and 3) debris aprons (LDA) or lineated valley fill (LVF) at lower elevations and on the valley floor. Further studies at MOC resolution of the surface of LVF/LDA deposits revealed that a large number of surface units are comprised of complex terrains consisting of pits and buttes. The occurrence of such units (also termed ‘brain terrain’; Noe Dobrea et al., 2007) across the surface of the LDA/LVF has been attributed to the significant loss of ice through sublimation (Mangold, 2003; Levy et al., 2009).

The three components of the fretted terrain environment characterized by Carr (2001) are clearly present within the LVF/LDA environments of the study region (Fig. 11, 12). Through the utilization of more recent data sets along with MOC images we have further investigated the surface textures present on these components within the study area. We find that distinctions between the main components can be made based on differences between their thermal properties in addition to the morphological differences reported by Carr (2001). Below we report on our findings.

Intermediate Units (IU) can be observed between the steep upper slopes and the rough textual (i.e. brain terrain) portions of both LDA and LVF deposits across the study region (e.g., see Fig. 11, 12). IU also comprise the surface of the LVF lobes emanating from the alcoves. The units are typically ~2 km wide and can extend continuously for up to several hundred kilometers along the base of plateau slopes and completely surround the flanks of mesas (Fig. 11). They appear as a region of relatively smooth terrain with respect to the rougher LDA/LVF surfaces at all spatial scales from HRSC (~ 18m/pixel) to HiRISE (sub meter) resolutions (Fig. 12). IU are also evident in THEMIS IR mosaics of the study area (Fig. 11), and appear relatively bright (higher temperature) compared to the adjacent LVF unit and plateau surface in both the daytime and nighttime data sets. In order to explain the higher nighttime temperatures we suggest that the IU have a higher thermal inertia than the surrounding units (Christensen et al., 2003). This in turn

is a proxy for grain size of the surficial deposits, suggesting that the IU consist of coarser material than that of the adjacent LVF (although it is worth noting that the reduction in sky exposure caused by the adjacent steep slopes may insulate the unit from radiative loss (Hecht, 2002), and thus contribute to higher nighttime temperatures). Nevertheless, the direct correlation between the temperature signal and the distinct morphology of the IU in the image data sets (Fig. 11) suggests that topography alone does not fully account for the higher nighttime temperatures.

Therefore, following Carr (2001), we interpret the IU to represent a relatively recent debris cover, the source of which is likely to be the adjacent higher exposed slopes (the steep upper slope unit of Carr, 2001). Boulders ~1m in diameter can be identified along the interior slopes of alcoves and the flanks of the main plateau (Fig. 13) forming scree which is interpreted to be the product of erosion along the slope. The slopes also appear very bright in the nighttime IR (Fig. 11), a finding that is consistent with other steep slopes elsewhere on Mars (Christensen et al, 2003). This is interpreted to be caused by the concentration of coarse material as a result of sub-area processes operating on slopes close to the angle of repose (Christensen et al., 2003). Evidence for mass wasting along the slopes in the study area exists in the form of spur and gully erosion along the plateau/alcove flanks that has produced highly linear troughs a few meters across (Fig. 13). We find no evidence for the involvement of liquid water in the formation of these features (as has been implied for larger gully forms; Malin and Edgett, 2000), and interpret them to represent critical slope failure and the generation of dry debris slides. The supply of such debris to the lower portion of the slopes could thus account for the formation of the IU. The LVF/LDA surfaces in contrast appear to be comprised of finer, reworked debris derived from sublimation of the underlying ice (Mangold, 2003; Levy et al; 2009). These may also be covered by a significant amount of eolian dust that has become trapped within the pits along the surface, which together could account for their lower thermal inertia.

Our findings suggest that the LDA/LVF deposits within the study area have undergone modification since their emplacement. This has been in the form of both potential ice loss from

sublimation related process to generate the pit and butte texture of their surfaces (which has been described by Mangold, 2003 and Levy et al., 2009) and through the formation of IU along their margins with the surrounding valley flanks. This demonstrates that for the majority of the surface of the LDA/LVF deposits, the original features associated with the last phase of activity are only preserved at longer wavelengths ($\sim <10$ m scale). Hence the current state of the LDA/LVF deposits is best described as the terminal remnant of previous activity when viscous flow last occurred (see Dickson et al; 2008).

5. Age Estimates of LVF/LDA Emplacement

Due to the wide distribution and integrated nature of the LDA/LVF deposits, their surfaces provide suitable areas on which to conduct crater size frequency distribution surveys and thus place constraints on the age of the deposits. The LVF/LDA deposits within the study region display impact craters with a similar range of degradation to that observed by Mangold (2003) within LDA deposits elsewhere across the northern dichotomy boundary. The study area displays crater morphologies which vary from fresh, circular depressions to flat ‘oyster shell’ and heavily subdued ‘ghost’ craters (Mangold, 2003). Mangold (2003) attributed the degradation of these craters to enhanced sublimation, which has also been responsible for the pitted texture present on the surface of LVF/LDA. More recent analysis of these small scale (< 1 km) craters on the surface of LVF/LDA deposits have found further evidence to suggest that impacts occurred into ice rich targets. McConnell et al (2006) used numerical and physical models to demonstrate the formation of central mounds due to ice rich targets. Kress et al (2008) noted a size distinction between smaller circular bowl-shaped craters and larger ‘ring mold’ craters, which they attribute to the difference between smaller impacts that occurred purely within a thin surface debris layer, and larger impacts that were large enough to pierce this upper layer and excavate ice-rich material below. The wide range in crater morphology attributed to impacts into ice rich targets and their

subsequent modification caused by enhanced sublimation means that crater counts will provide a minimum age for the emplacement of the LDA/LVF units.

The largest continuous extent of LVF/LDA in the study region is present within the extensive elongated fretted valley to the south of the main plateau. Due to the distinctions between the intermediate units and the surface of the LVF deposits that were discussed earlier, we focused our crater counting on the LVF units that provide a continuous area extending for ~190 km in the main southern valley. All identifiable craters larger than 100 m were counted regardless of their morphology or apparent level of degradation. One HRSC image (orbit 1589) was used to maintain a consistent resolution and image quality over the entire count area. Isochrons plotted according to the Hartmann (2005) system were used and compared to our results (Fig. 14). These show good agreement with the 100 Ma isochron, though slightly offset to the right for craters larger than 250 m. This indicates a surface age for the LVF within the Late Amazonian (>100 Ma – 500 Ma), a result that is consistent with other published results of LVF/LDA surface ages (e.g., Mangold, 2003). For craters smaller than 250 m, there is a noticeable downturn in the crater distribution, suggesting a period of resurfacing. This further supports the occurrence of mantling and degradation due to sublimation operating along the surface of the LVF/LDA deposits (Levy et al., 2009). The absence of the smallest craters following any of the isochrons indicates that the degradation process may still be active or at least was active until very recently (see also Mangold, 2003), which is consistent with the current instability of ice on the surface of Mars at these latitudes.

6. Evidence for the Previous Extent and Thickness of LVF

The acquisition of very high-resolution images (MOC, CTX, HiRISE) provides an opportunity to search for evidence for the previous distribution of LDA and LVF (on alcove walls, as high stands and trimlines, and evidence for ice on adjacent plateaus, etc). The instability

of water ice throughout the martian year at latitudes where LVF deposits at present (Farmer and Doms., 1979; Mellon and Jakosky., 1995) suggests that volatile-rich landforms at these latitudes will lose ice to the atmosphere over time by vapor diffusion and sublimation. As was previously discussed, evidence for significant sublimation is apparent on the surface of LVF deposits, as they exhibit a pit and butte/brain terrain texture, the generation of which has been attributed to substantial sublimation (Mangold., 2003; Levy et al., 2009).

The debris-covered glacial model for LVF formation interprets the presence of the debris cover as resulting from the production of a surface lag formed on the deposits by the concentration of debris material through the sublimation of ice. This process results in a sublimation till, similar to that seen on debris-covered glaciers in the Antarctic Dry Valleys (Marchant et al., 2003; Marchant and Head, 2007). The formation of the sublimation till serves to inhibit further loss of ice (Head et al., 2006a). In addition, during the transition from active glaciation to the currently observed state of preservation of the deposits, ice in the accumulation zones will change balance from net annual ice accumulation to net annual ice loss. This means that glacial flow will continue for some time after the change in this critical balance, but that the glacial system will undergo both lateral retreat and vertical downwasting. By extension, this suggests that significant ice has been lost from LVF, and indeed that entire glaciers might have been lost in areas where the debris supply was limited and thus prevented the formation of protective sublimation tills. Utilizing HiRISE data, Dickson et al (2008) found evidence for the occurrence of linear ridges that run along the internal walls of alcoves that hosted individual lobes of LVF. These features were interpreted to be lateral moraines that were deposited when the LVF deposits were at least 900 meters higher than at present. We also investigated the available HiRISE coverage of the study region in order to search for the presence of similar landforms.

The highest resolution views (25 cm/pixel) of the flanks of the main plateau reveal linear features that are orientated parallel to the slope contours (Fig. 15). These linear features are ~10 m wide and extend (sometimes discontinuously) for 100s of meters. Shadows cast by these

lineations indicate that they are positive features. The ridges are too small to be resolved in either MOLA or HRSC DTMs, although the shadows they cast indicate that they are of the order of several meters high. The ridge-like nature of these landforms argues against an origin as exposed bedrock and suggest that they may be depositional features. Boulders (about a meter in diameter) are present on the flanks of the ridges and form scree (Fig. 15), which may have originated from the degradation of the ridges. If this interpretation is correct, it further suggests that the ridges are comprised of largely unconsolidated material, including boulder-sized debris.

On the basis of the morphology and orientation of these features, we interpret them to be lateral moraines that were deposited along the margin of the ice when the LVF material was at a higher elevation, and occupied a greater portion of the valleys and alcoves. The difference in elevation between the current surface of the LVF and the highest ridge is 800 m, thus suggesting that at least this amount of ice has been lost from the current LVF deposits. This is consistent with the findings of Dickson et al (2008), who showed evidence for loss of over 900 m of ice in LVF along the dichotomy boundary. Terrestrial lateral moraines require a supply of debris that is usually produced by the mass wasting of slope material above the ice (similar to what was observed and discussed in section 4.3, see Fig. 13), or transported laterally along the margins from debris forming in the accumulation zone. Therefore, if this interpretation is correct, this observation provides a minimum estimate of the previous thickness of the LVF deposits, as ice above this elevation would have been above sources of debris that are required to form the lateral moraines.

7. Evidence of Multiple Phases of LVF Emplacement

Within the overall pattern of integration present in the LVF and LDA systems in the study region (Fig. 5), there is a distinct group of small lobate flow units that appear at THEMIS resolution to be superimposed on, rather than coalescing with, the surrounding larger scale

LVF/LDA systems (Figs. 9, 16). These flow units are located along the southwestern flanks of the main plateau and are associated with individual source alcoves; they appear similar in both scale and morphology to the LVF lobes that have been identified emerging from other alcoves along the main plateau. As was observed with the LVF lobes (Fig. 7-10), the superimposed flow units display surface lineations, including longitudinal ridges that are orientated along the center of the deposit in the downslope direction (Fig. 16), producing patterns of convergence between topographic obstacles indicative of terrestrial debris-covered glacial flow (Head et al, 2006b). The flow units terminate downslope of the mouths of their source alcoves in the form of expanded lobes, further supporting the argument that these features are comprised of ice-rich materials. In contrast to the other LVF lobe features (Fig. 7 – 10), the snouts of the superimposed units remain completely distinct from the main LDA bodies onto which they emerge, with no evidence of merging, blending and integration at the point of contact between the two units, or any other form of disturbance within the main LDA body downslope of the contact area (e.g. compare Fig. 9 with 16).

This suggests that the superimposed features are stratigraphically separate units that overlie, and are thus younger than the surrounding main trunk LDA systems. If this is indeed the case, it implies that the units were emplaced and active after the cessation of flow within the main integrated bodies LVF/LDA. Similar features have also been identified within the LVF systems present in both Nilosyrtris Mensae (Levy et al, 2007) and the Protonilus Mensae-Coloe Fossae region (Dickson et al., 2008) of the dichotomy boundary. These features have been termed ‘*Superposed LVF*’ and ‘*Small-scale Superposed LVF*’ by Levy et al, (2007), with the distinction being the scale of the feature. The *Small-scale Superposed LVF* is described as being present in only small-scale valleys of the order of a kilometer wide by five kilometers long and terminate in abrupt convex-up lobate fronts that are distinctive from and lie above main trunk bodies of LVF. Levy et al (2007) attribute these flows to alcove microclimates that permitted the accumulation of small volumes of ice during periods of climatic conditions unfavorable to large scale regional

glaciation. Within the Protonilus Mensae-Coloe Fossae region there are examples of superimposed lobes that exhibit terminal moraines that sit above the main trunk valley LVF (Dickson et al, 2008). The lack of modification and/or deflection within the superimposed moraines is highlighted by Dickson et al (2008) as evidence that the lobe does not represent a reorganization of ice flow during the waning stages of glaciation, but rather was emplaced during a later renewed phase of glacial activity.

The occurrence of identical superimposed landforms thousands of kilometers apart suggests that additional features may exist elsewhere across the dichotomy boundary (Fig. 17), and should be a focus of attention for future research. The distance between the locations in which these features have formed suggests that a more recent phase of glaciation has occurred across the northern dichotomy boundary since the emplacement of the main bodies of LVF/LDA around 100 – 500 million years ago. The small size (1-2 km wide) of the more recently emplaced LVF lobes argues that if the glacial model for LVF/LDA formation is correct, then the most recent glacial phases must have been characterized by climatic changes of a more limited magnitude and shorter duration.

8. Conclusions

We used new spacecraft data to carry out an in-depth investigation of lineated valley fill and lobate debris apron systems in a previously undocumented area of the dichotomy boundary. Our investigations provide persuasive evidence that debris-covered glaciation has played a significant role in the formation of these deposits. We reach the following conclusions:

- 1) Evidence has been documented for numerous, localized alcoves, enclaves and tributary valleys within the walls of plateaus and mesas which provide a source of small scale LVF flows and lobes; we interpret these flows to be the remnants of debris-covered glaciers that formed from the deposition of ice within the alcoves during previous preferential climatic regimes associated

with higher planetary obliquities (Madeleine, et al., 2007);

2) We found abundant evidence that LVF and LDA within the study region is comprised of the same material, show integrated flow patterns, and are of the same origin; this is supported by both landform types being supplied from individual lobes from within alcoves and the observations that LVF and LDA integrate seamlessly throughout the study area;

3) We present evidence to suggest that the LDA/LVF deposits within the study area have undergone a significant amount of modification since their emplacement. This has been in the form of both potential ice loss from sublimation related process to generate the pitted and butte texture of their surfaces (which has been described by Mangold, 2003 and Levy et al., 2009) and through the formation of IU along the margins of the LVF/LDA with the surrounding valley flanks due to recent mass wasting processes.

4) We have mapped out large scale integrated systems of LVF and LDA, reminiscent of terrestrial glacial land systems; the nature of the integration is well illustrated by the expressions on the surface of LVF displaying compression between topographic obstacles and divergence around them;

5) We present evidence for former highstands of the LDA/LVF that suggest that the surface may have been more than 800 meters higher during the full glacial phase than the thickness of the residual deposits today;

6) On the basis of superimposed impact craters, we derive a late Amazonian age for emplacement of the main body of LVF/LDA deposits. This may have coincided with a period of high obliquity, causing the loss of water ice at the polar caps and the redistribution of volatiles to lower latitudes (Madeleine et al., 2007).

7) We outline evidence for multiple phases or cycles of LVF emplacement, possibly due to further perturbations of orbital parameters that were of a lesser extent than that responsible for the original LDA/LVF emplacement;

8) Modification of the fretted valleys of the dichotomy boundary has been substantial

locally, as evidenced by the erosion of numerous alcoves, several – tens of kilometers in length into the flanks of the highlands and isolated mesas. However, we find no evidence that the Amazonian glacial epochs caused substantial retreat of the dichotomy boundary measured in the tens to hundreds of kilometers.

Our findings support the results of an analysis just to the east of the study region (Head et al, 2006b) and of studies carried out elsewhere along the dichotomy boundary (Head et al, 2006a; Levy et al., 2007; Dickson et al., 2008; Kress et al., 2008) that find further evidence for the past presence of debris-covered glaciers and extensive valley glacial landsystems.

Acknowledgments

We gratefully acknowledge financial support from the NASA Mars Data Analysis program (NNX07AN95G), the NASA US participation in the Mars Express High-Resolution Stereo Camera (HRSC) (JPL1237163), and the NASA Applied Information Systems Research Program (NNG05GA61G). Thanks are also extended to Caleb Fassett and James Dickson for their technical assistance in data analysis and contribution to scientific discussions during preparation of the manuscript.

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Figure Captions

Fig. 1. Context MOLA topographic map of the northern dichotomy boundary at Deuteronilus-Protonilus Mensae; the location of the study region is highlighted by the box. Note the gradual decrease in topography from the southern highlands into the northern lowlands and the abrupt scarp. The boundary area is characterized by the occurrence of numerous mesas that become increasingly smaller to the north.

Fig. 2. a) Oblique view (facing directly to the north) of the study region represented in a

MOLA DEM. We focus on the plateau in the upper right portion of the image containing the 60 km diameter Sinton impact crater and the adjacent valleys. At MOLA resolution, Lineated Valley Fill (LVF) appears as a smooth deposit and is observable throughout the valleys surrounding the main plateau. b) Map of the study area highlighting the extent of the LVF and Lobate Debris Aprons (LDA) systems within the region. The dashed white line indicates the area containing the LVF and LDA deposits, which are found above an elevation of -3200 m. Boxes indicate the locations of images in Figs. 4 – 7 and 5. The dashed lines correspond to topographic profiles in Fig. 4. The map consists of a portion of HRSC image 1589, overlain by MOLA topographic data.

Fig. 3. Slope map. The fretted terrain within the study area consists of relatively smoothed surfaced plateaus and mesas surrounded by steep flanks exhibiting slopes of over 23° . Smooth crater-fill deposits line the floor of Sinton crater and the terraces along its interior flanks. The LVF/LDA deposits surrounding the main plateau exhibit almost entirely horizontal surfaces (slopes of less than 0.5° , blue shades), but are made identifiable by their convex termini which have slopes $3.5 - >5^\circ$ (orange - red shades). Such morphology is consistent with other LVF/LVA studies (e.g. Mangold, 2003; Head et al, 2006b) and with terrestrial glacial profiles. The map was derived from the processing of MOLA data using Arc GIS software and was overlain on a portion of HRSC image 1589.

Fig. 4. Topographic profiles extending both N-S and W-E across the study region, derived from MOLA 128 pixels/degree gridded data set. The profiles demonstrate the lobate nature of the termini of both LVF and LDA deposits that are reminiscent of the profiles of terrestrial glaciers. The main plateau exhibits a similar surface elevation to the surrounding mesas (profiles B and C) despite the disparity in surface area between them. Such profiles are consistent with glacial erosion alongside mesas, reducing their overall volume but leaving their summit elevations intact; over time this has resulted in the gradual ‘mesa-ization’ (Head et al, 2006b) of

the northern margin of the dichotomy boundary.

Fig. 5. Geographical representation of the general flow trends observable within the LVF systems along the western flanks of the study region, compiled using HRSC, THEMIS and MOC images. a) Viking composite of the study area, which clearly shows the full extent and integrated nature of the LVF/LDA systems present. b) Flow trends of LVF/LDA. Alcove lobate flow-like features are represented by small arrows, with thicker arrows depicting more broad flow trends of LVF on the valley floor. The integrated systems are similar to those mapped by Head et al (2006b) to the east of the study region. Boxes show the location of images in Fig 6. Base map is MOLA gridded topography, 100 m contour interval.

Fig. 6. Examples of patterns of divergence and convergence visible within the surface of the LVF systems mapped in Fig. 5. a) Convergence of LDA between a gap within a topographic ridge (large arrow) forming a 3 km wide lobe of material on the downslope side. The main branch of LDA is fed by small lobes (small arrows) emanating from several kilometer wide alcoves along the northern flanks of the plateau. THEMIS image V12506004. b) Convergence and integration of flow between a 10 km wide mesa and the main plateau (highlighted by large arrow). Small lobes can be seen flowing from the flanks of the mesa (small arrows). THEMIS image V03282003. c) Divergence of flow around a 5 km mesa. The flow originates from small lobes within alcoves to the west of the image (small arrows) and flows out the north and east around isolated mesas. CTX image P13_006239_2240.

Fig. 7. Integrated LVF system within a complex alcove comprised of multiple branching tributaries. (a) The alcove extends 20 km into the plateau, almost breaching it at its furthest point (X) and consists of four major tributaries, all of which are filled with LVF. Lineations on the surface of the LVF are aligned in the downslope direction within each of the tributaries (shown

by arrows in b) and converge together at the mouth of the alcove. A large valley system is observable in the southeastern section of the image and appears to connect to the easternmost tributary of the alcove, though its mouth is superimposed by the fresh ejecta surrounding the crater at point Y. Small lobes can be seen emerging from indentations within the northern wall of the main plateau and integrating with the main body of LDA, demonstrating how large-scale LDAs are fed by individually sourced lobes. (a) Portion of HRSC image 1589.

Fig. 8. Folds and other features providing evidence of viscous flow on the surface of a LVF lobe(a). (b) Dashed lines indicate the location of undulating ridges along the surface of the lobe and are interpreted to be related to the downslope flow of ice-rich material along the alcove. These ridges show evidence of folding in the downslope portions of the alcove due to compressional forces produced by the lobe flowing into the main trunk of LVF within the southern valley. MOC image R0700879.

Fig. 9. Relationships between lineated valley fill within a complex alcove and the large-scale lobate debris apron that it feeds. (a) Individual lobes of lineated valley fill emerge from local enclaves at the end of, and along the margins of, the branching alcove (shown by short arrows in b). These converge in the central parts of the branching alcove in the form of parallel ridges and troughs orientated parallel to the valley walls, which become a single lobe of lineated valley fill. Having passed through the narrow restriction at the edge of the main trunk valley, the LVF expands out into a distinctive dual-lobe pattern. Ridges, presumably formed by subsequent flow events out of the complex alcove (bottom center, this portion of the image has been stretched for clarity) can be seen to bend toward the left and begin to merge with the broader patterns of the LDA on the floor of the trunk valley. Thus, numerous LVF sources in the branching alcove tributaries are seen to converge and contribute to the branching alcove flow that makes up the LVF, and then collectively contribute to the main trunk of LDA. Patterns of ridges

in these examples are very similar to those seen in cold-based, debris-covered glaciers forming in branching valleys in the McMurdo Dry Valleys, Antarctica (c). Mullins valley, a tributary valley of Central Beacon Valley, shows concentric ridges produced by compression as material moves down-valley toward the valley mouth before turning to join the trunk valley. Within the image there is also a small scale lobe that appears to be superimposed on the surrounding LVF rather than merge with it in a similar fashion to the lobe with in Fig. 16, and thus may represent a later-stage accumulation of ice and subsequent viscous flow. The box present within a) provides the location of the image in Fig. 13. (a) Portion of THEMIS image V11620008. (c) TMA 3080 Frame 276 taken November 21, 1993, part of the Taylor Valley LTER series.

Fig. 10. Direction of flow of material within the main trunk valley LVF deposits inferred from the integration with LVF material emanating from an alcove within the southern portion of the main valley. At the point where the lobe merges with the main body of the LVF, the lineations along the surface curve to the southeast, despite the fact that the alcove is oriented in a southwesterly direction. This indicates that at the time that the LVF deposits were formed, the main trunk of LVF within the southern valley flowed in an easterly direction. This demonstrates how LVF accumulating within an alcove could have supplied main trunk valleys with ice-rich material when the deposits were active. At this spatial resolution (3 m/pixel) such lineations appear as accumulations of ridges and knobs (depicted as horizontal lines in b) that are interpreted to be the result of sublimation of ice from within the LVF deposit (e.g. Mangold, 2003). MOC image E1900479.

Fig. 11. a) HRSC image of the study area highlighting the IU. b) Mosaic of Nighttime THEMIS infrared images of the southern portion of the plateau. The mosaic has been normalized to account for the different times of night at which the individual images were taken. Intermediate Units (IU) present along the flanks of the plateau and within the alcoves themselves can be seen

as regions of brighter pixels corresponding to relatively higher temperatures, which in turn is most likely correlated with a relatively high thermal inertia. Thus we interpret the IU to represent a debris cover of relatively coarse material compared to that which comprises the surface of the main body of LVF. (a) The location of figures 12 and 13 are highlighted. Image is a subset of HRSC 1589.

Fig. 12. MOC image of three main components of the fretted terrain environment identified by Carr (2001): 1) a steep upper slope, where bedrock may be visible, 2) intermediate units (IU) which appear smooth, and 3) LDA or LVF (which in the case of this image is LVF). The surface of the LVF is comprised of pits and buttes. The context for the figure is provided in figure 11. MOC image: R1201629.

Fig. 13. High-resolution ($\sim 0.3\text{m/pixel}$) view of the main plateau flanks adjacent to an alcove containing a source lobe of LVF (see Fig. 9). Scree, containing boulders (meters in diameter), is resolvable along the plateau flanks and is a plausible supply of debris to the IU along the surface of LVF/LDA deposits surrounding the plateau. Erosional features in the form of spur and gully morphology are testimony to the occurrence of mass wasting along the slopes. The context for the image is displayed in Fig. 9, a) and Fig. 11. Image is a subset of HiRISE image PSP_006806_2215.

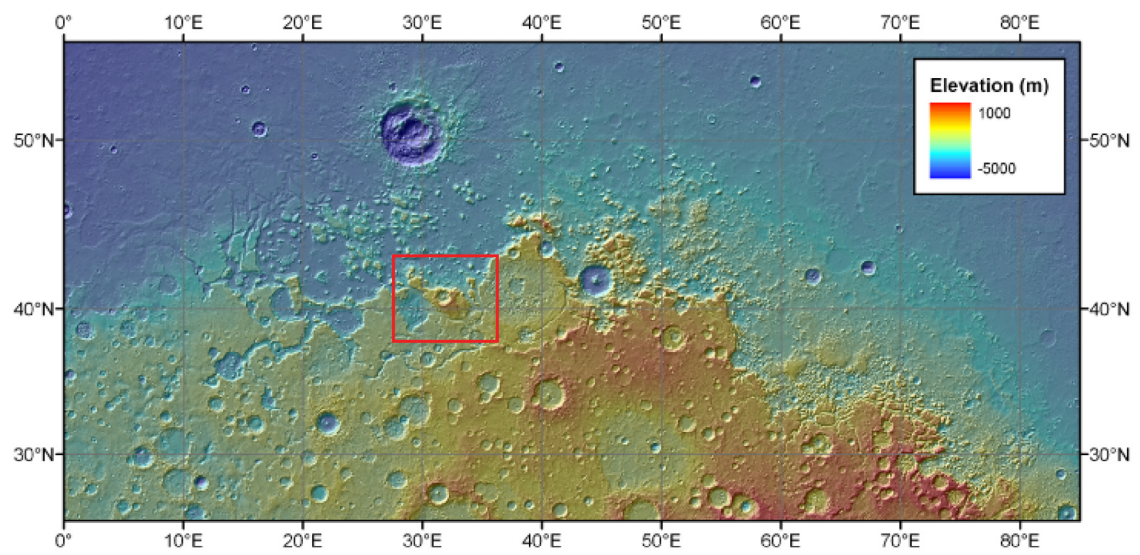
Fig. 14. Results of crater counts conducted along the surface of the main trunk of LVF within the valley to the south of the main plateau. The results are plotted against isochrons according to the Hartmann (2005) system. The plot suggests an age of 100 – 500 Ma for craters larger than 250m.

Fig. 15. Linear ridges along the walls of alcoves cut within the flanks of the main plateau.

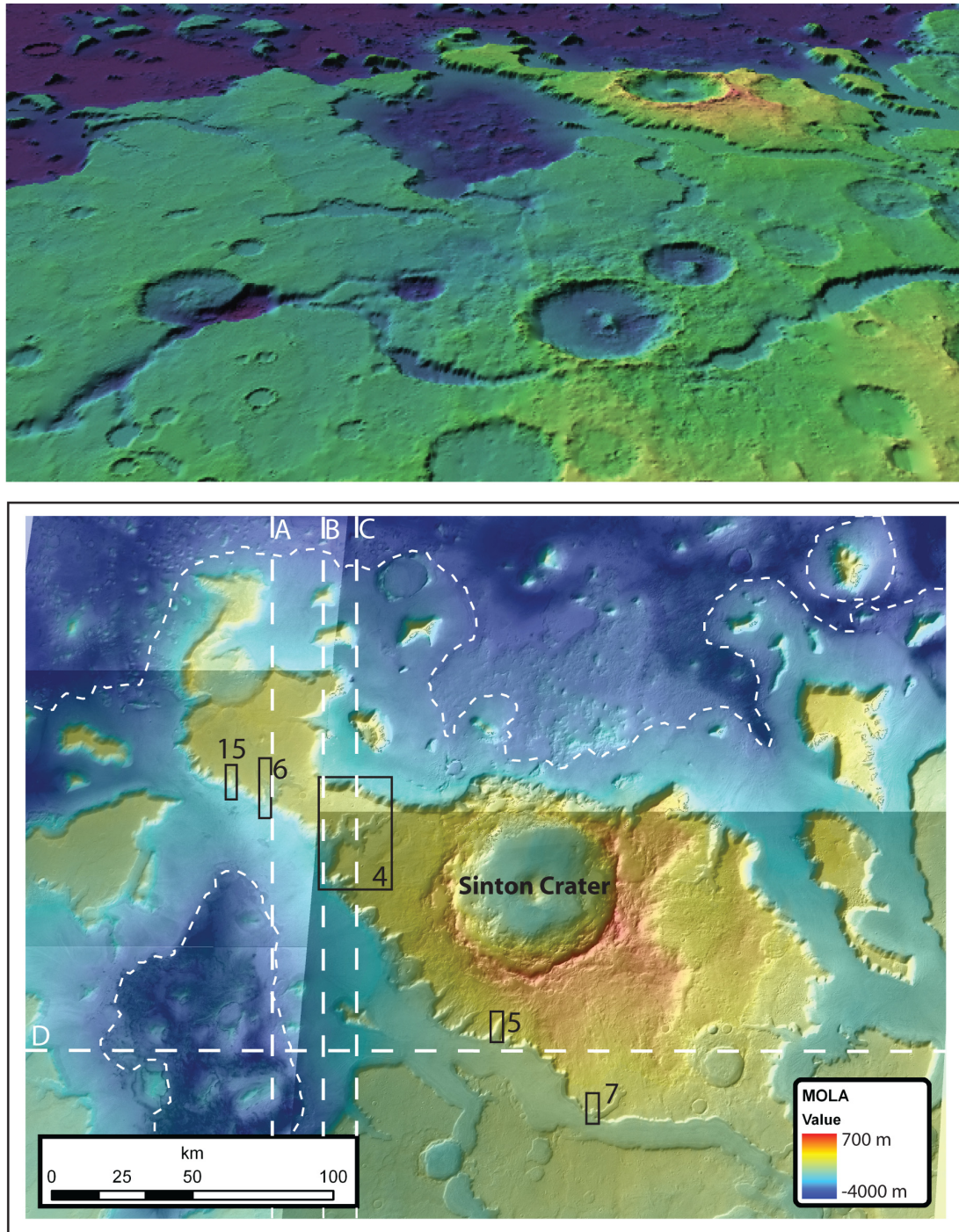
We interpret these ridges to be lateral moraines that represent the former elevations of LVF. (Top image) Context view of region demonstrating the parallel nature of the ridges that are located at all elevations above the current locations of LVF deposits and extend for 100s of meters in length. (Lower left image) Close-up view of the edge of the alcove. The ridges form discontinuous patterns and display convex forms in the downslope direction. (Lower right image) Highest resolution views of the ridges demonstrating their topographic form, which is highlighted by the occurrence of shadows to the north of the ridges (sun from the southwest). Boulders form scree on the slopes extending up to the ridges. All images are subset of HiRISE image: PSP__006806_2215, north is to the top of the images.

Fig. 16. Lobe of LVF emerging from an alcove that appears to be superimposed on, rather than merging with, the main branch of LDA onto which it flows (left panel). The entirety of the lobe is visible above the main body of the LVF, and there appears to be none of the disturbance within the main body such as is observable in Figs. 9 and 10. (b) The direction of flow of the lobe is indicated by the small arrows; the large arrows indicate the flow patterns of the main branch of LDA onto which the lobe sits (right panel). Other such features have been observed and documented in other regions of the dichotomy boundary (e.g. Levy and Head., 2007; Dickson et al, 2008), and are interpreted to represent a second, later and less extensive phase of glacial emplacement. HRSC image 1589.

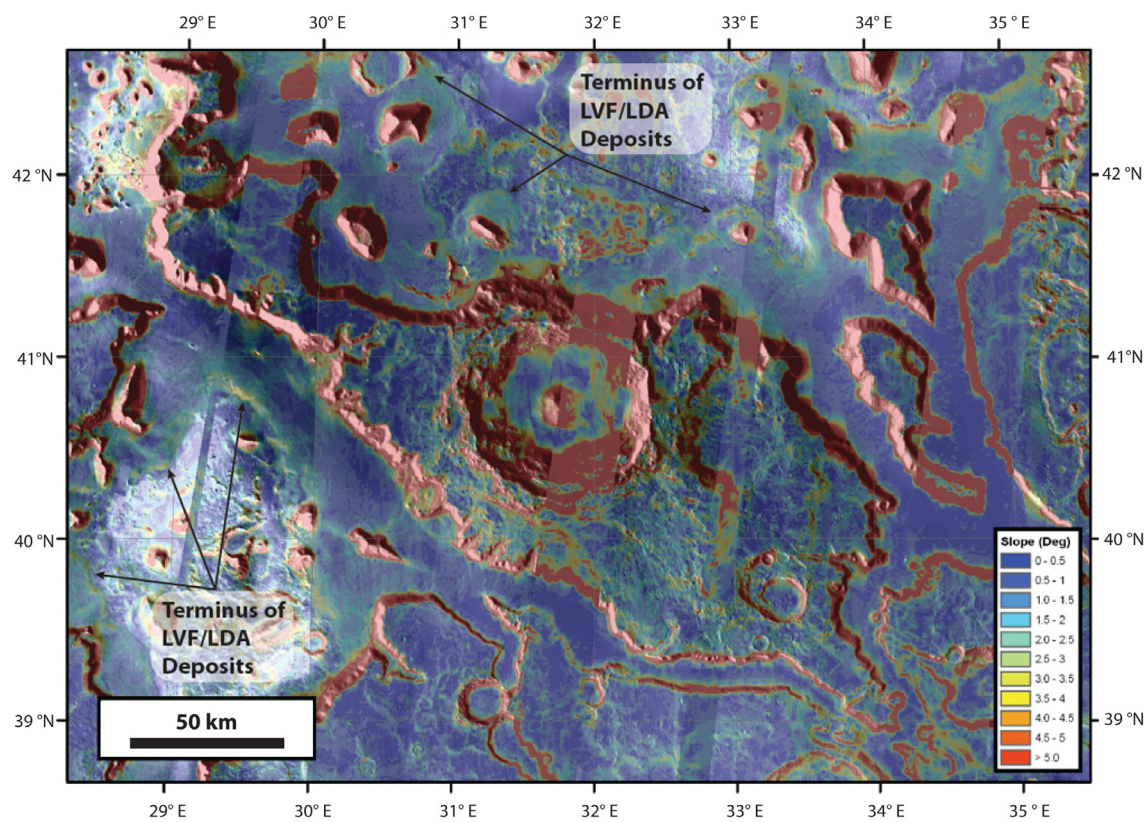
Fig. 17. The locations of detailed studies that have identified evidence for multiple episodes of LVF emplacement across the northern dichotomy boundary. The occurrence of such features over 2000 km apart suggests that more recent episodes of LVF activity were occurring across the northern dichotomy boundary. Map is a shaded relief of gridded MOLA topography data.



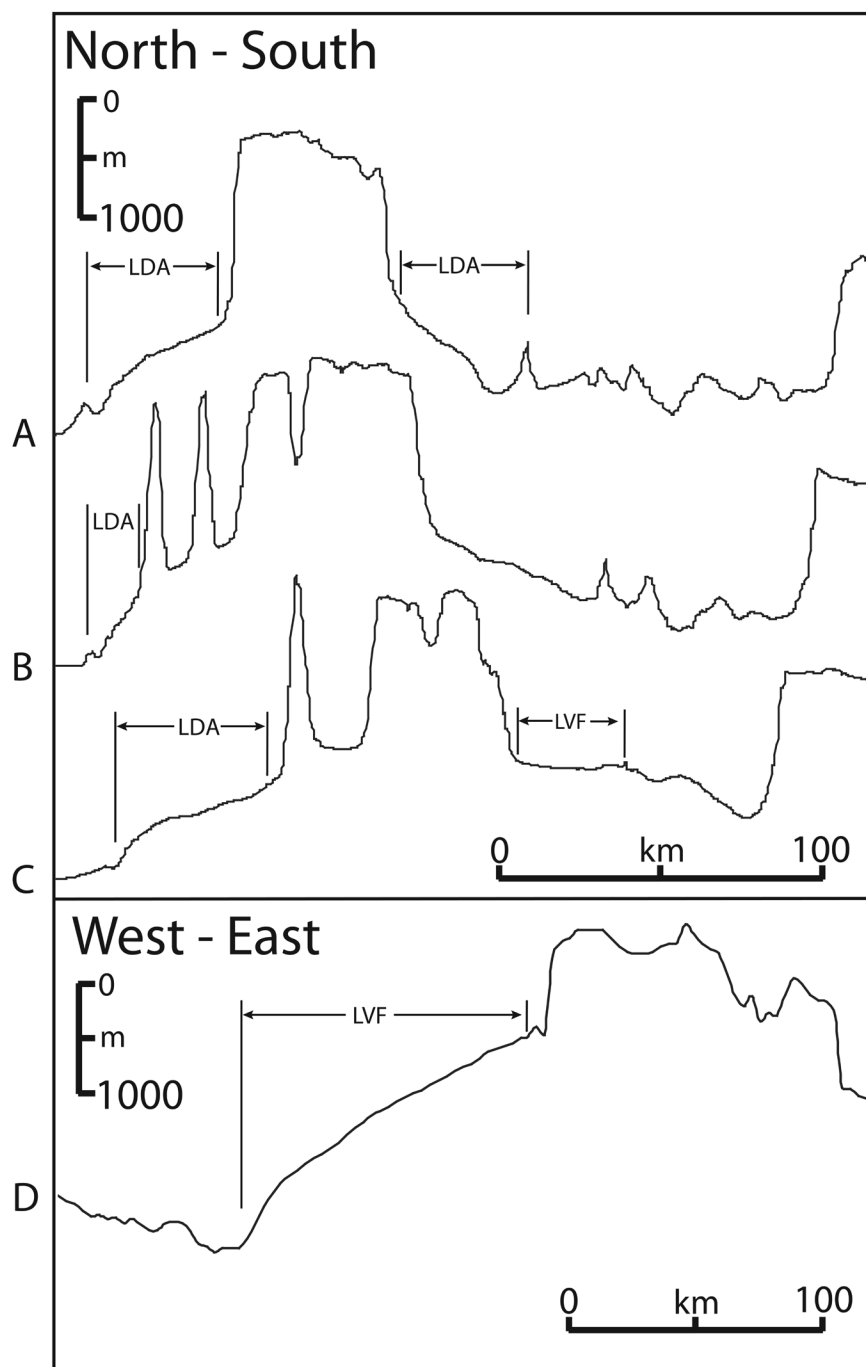
Chapter 2, Figure. 1



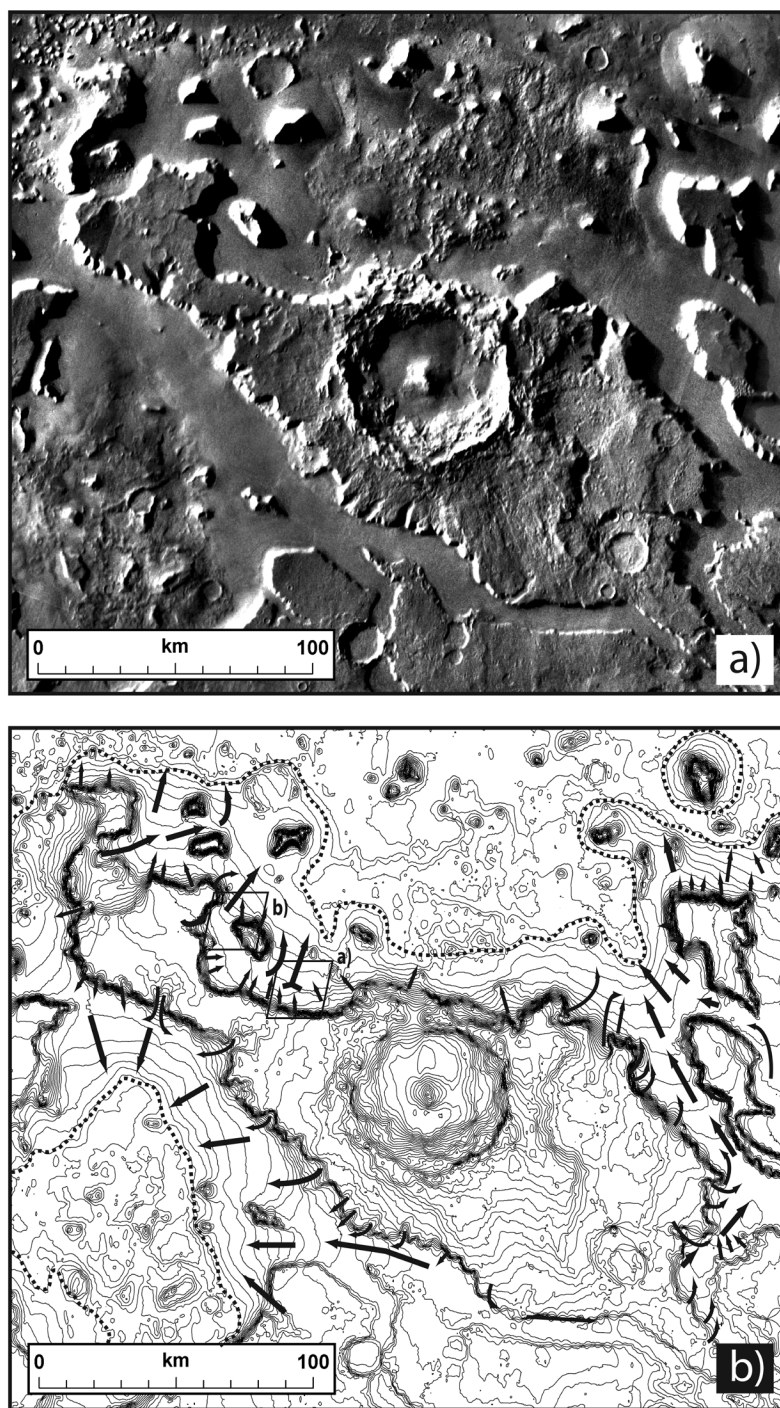
Chapter 2, Figure. 2



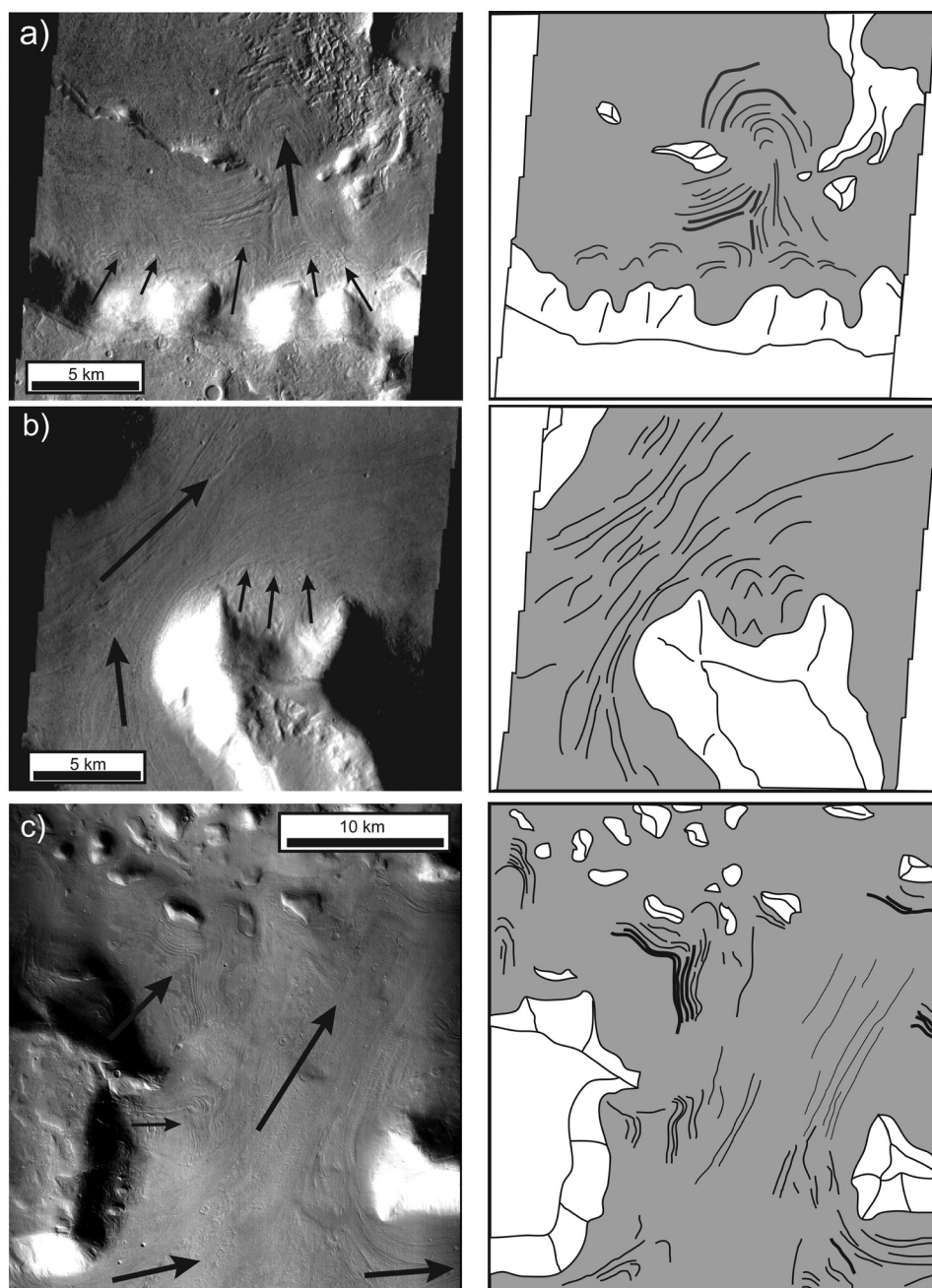
Chapter 2, Figure. 3



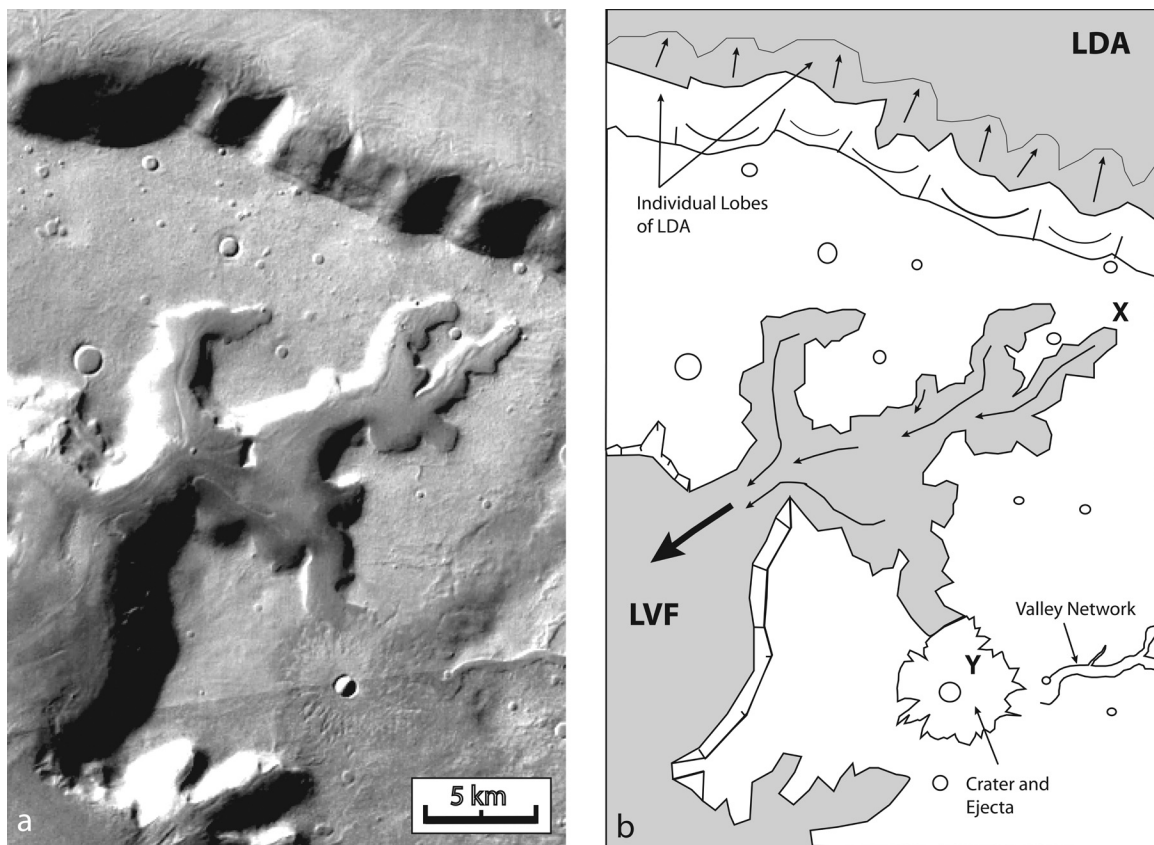
Chapter 2, Figure. 4



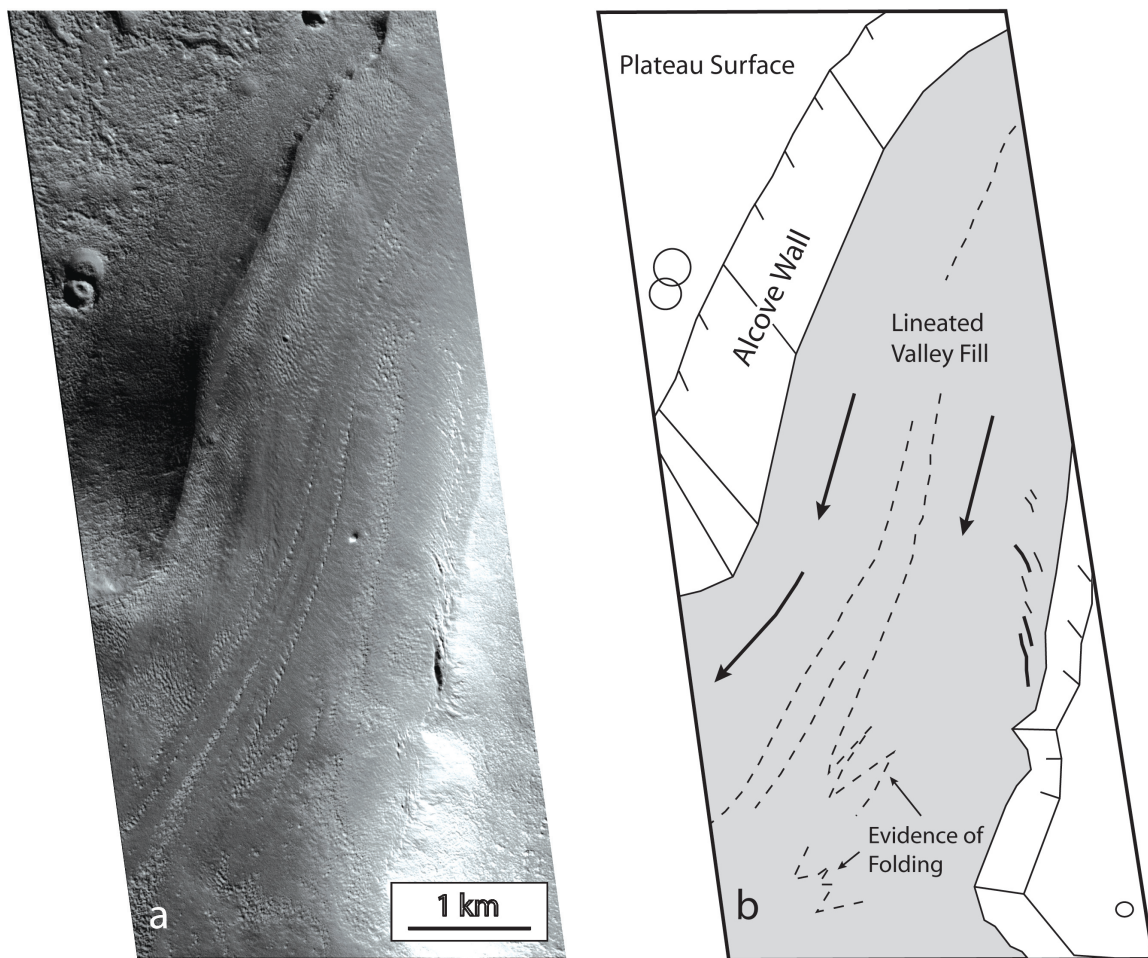
Chapter 2, Figure. 5



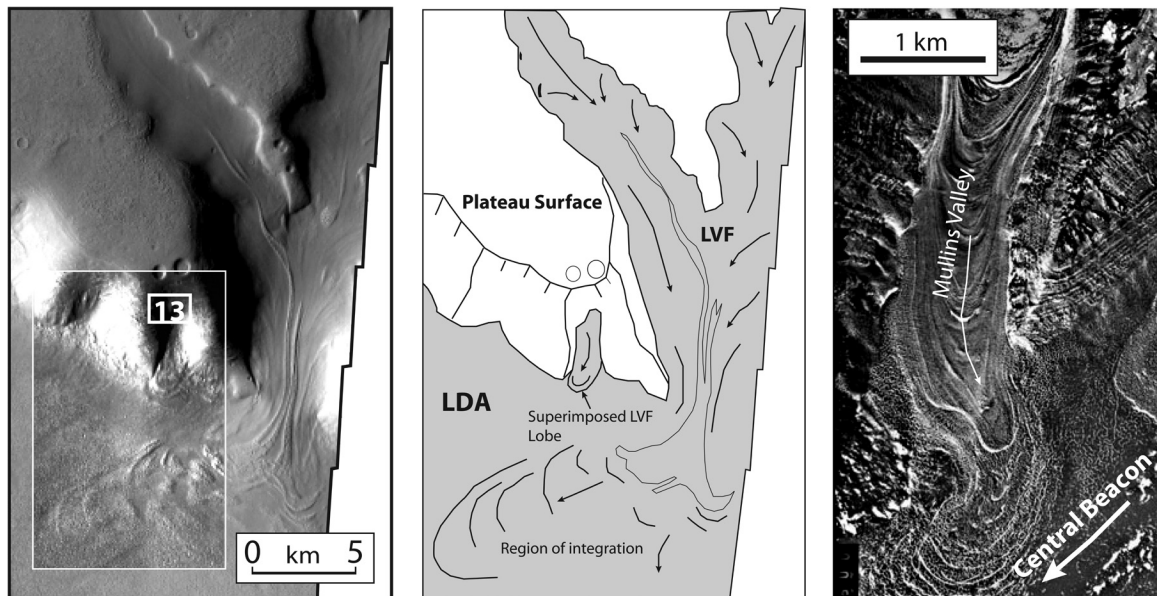
Chapter 2, Figure. 6



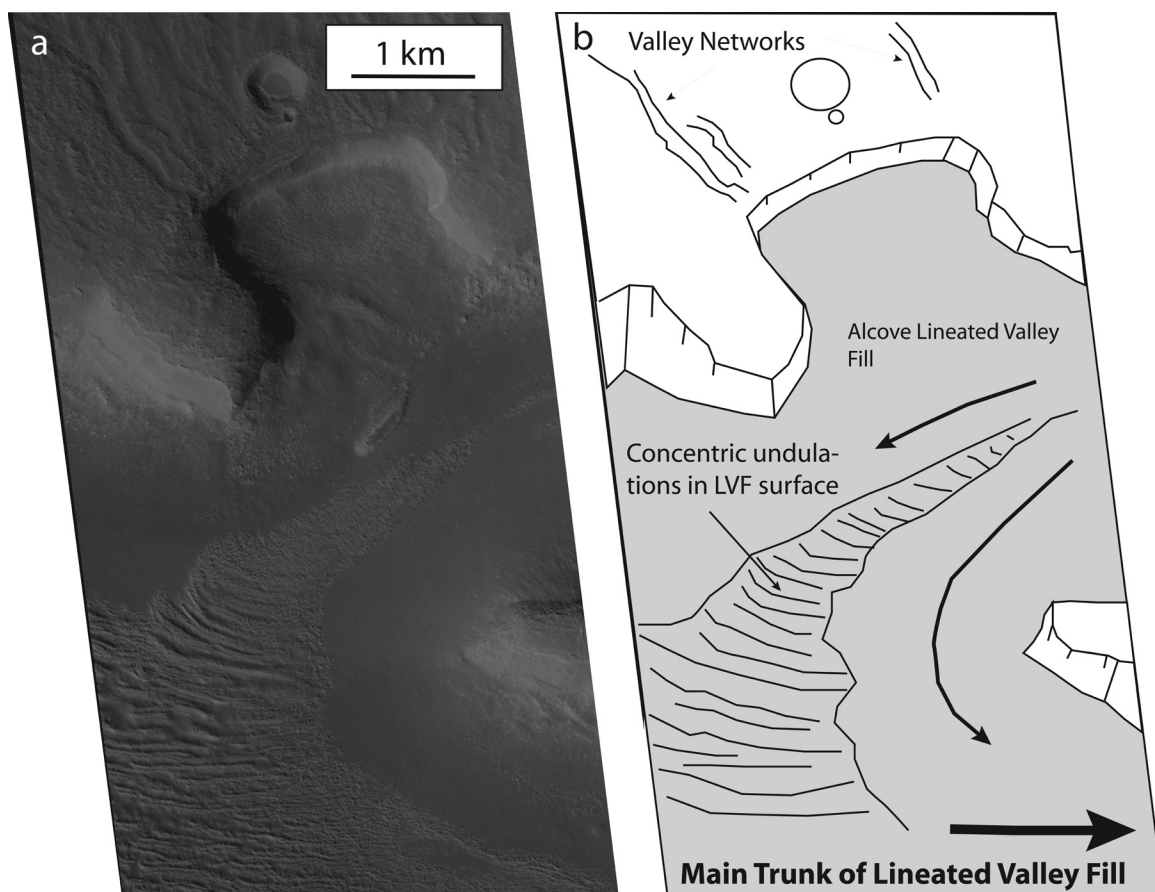
Chapter 2, Figure. 7



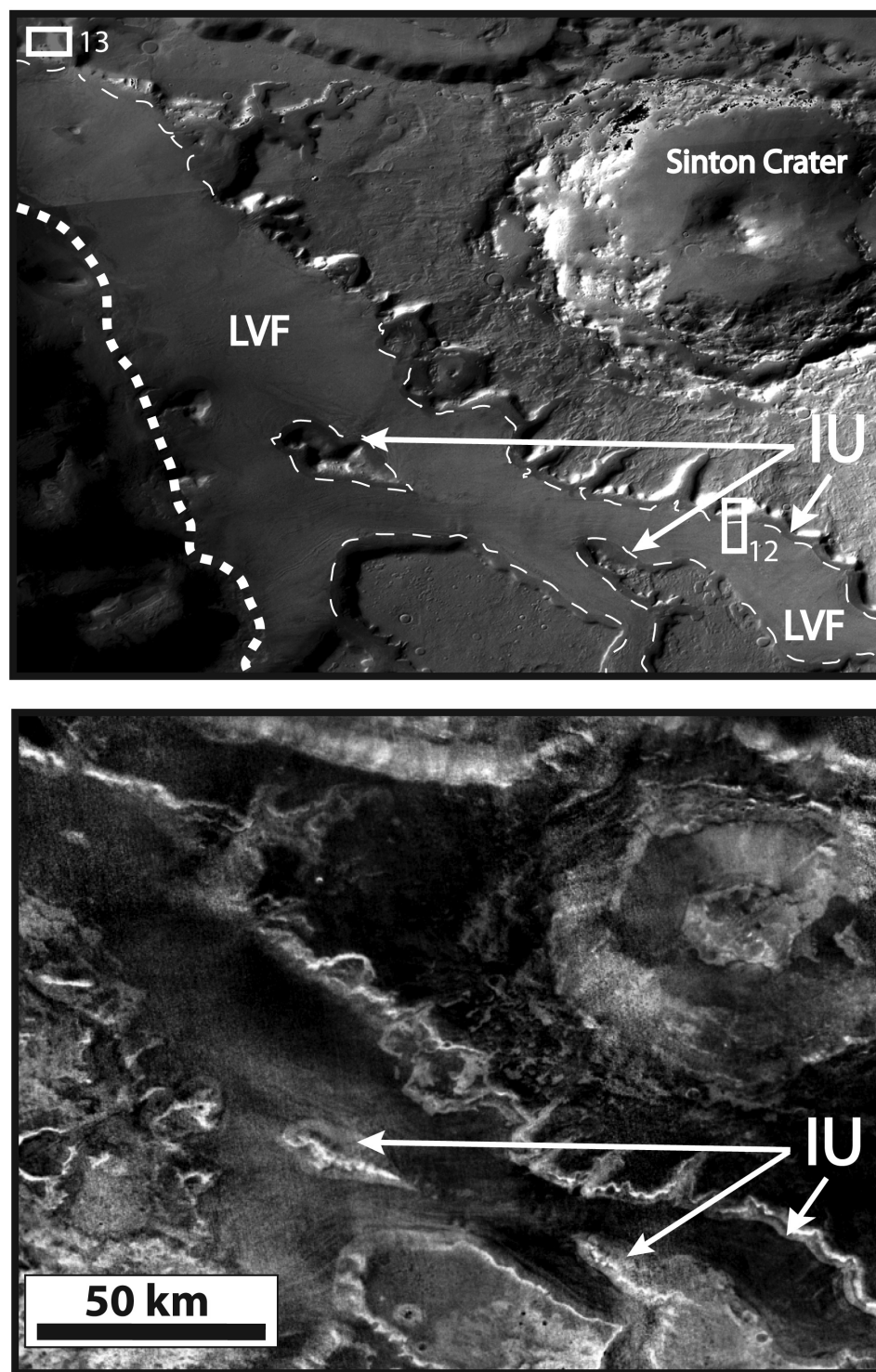
Chapter 2, Figure. 8



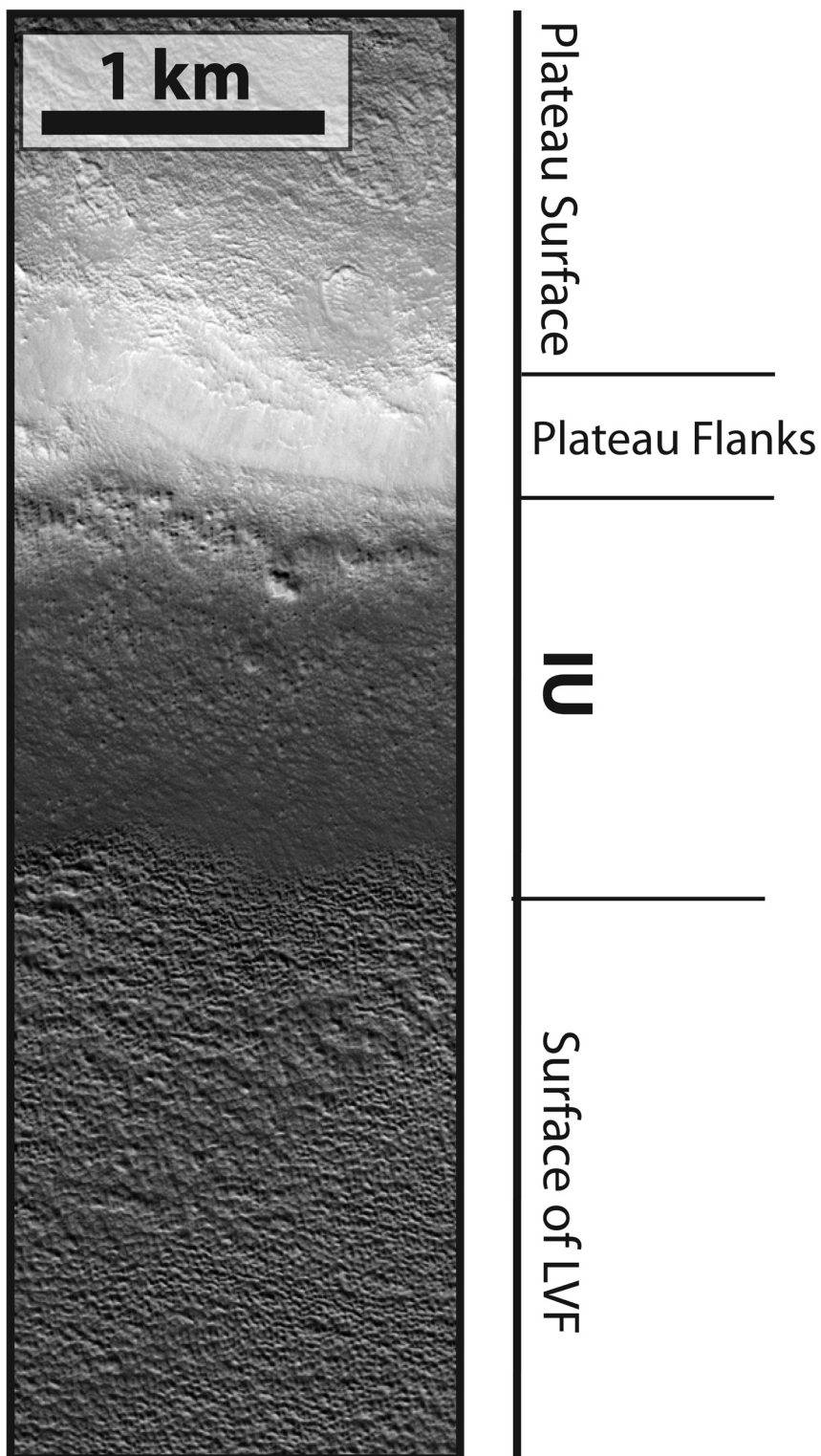
Chapter 2, Figure. 9



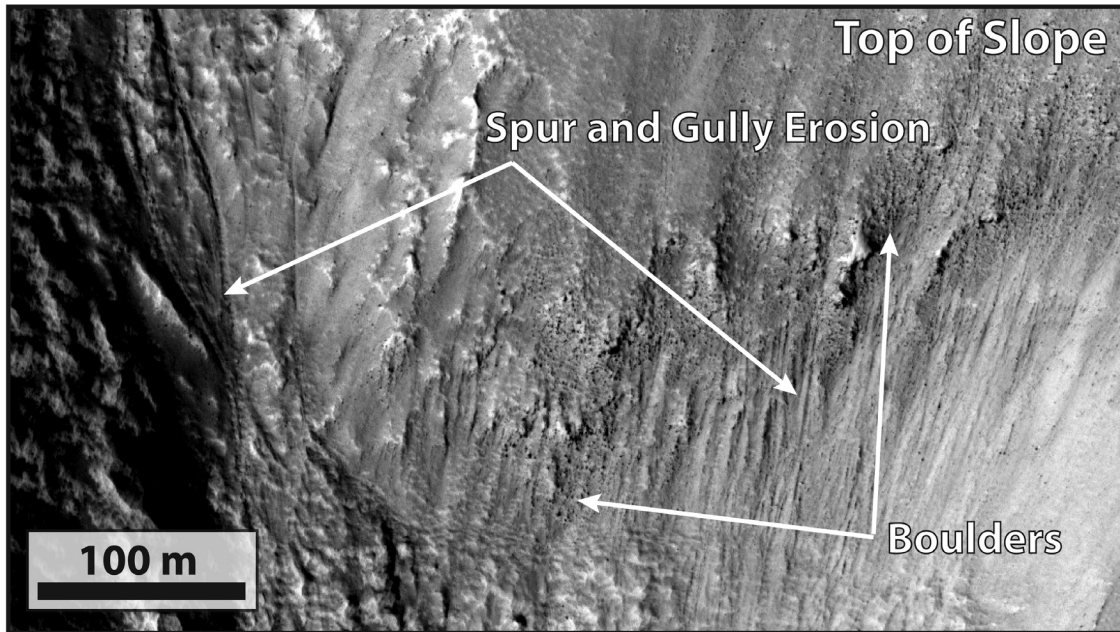
Chapter 2, Figure. 10



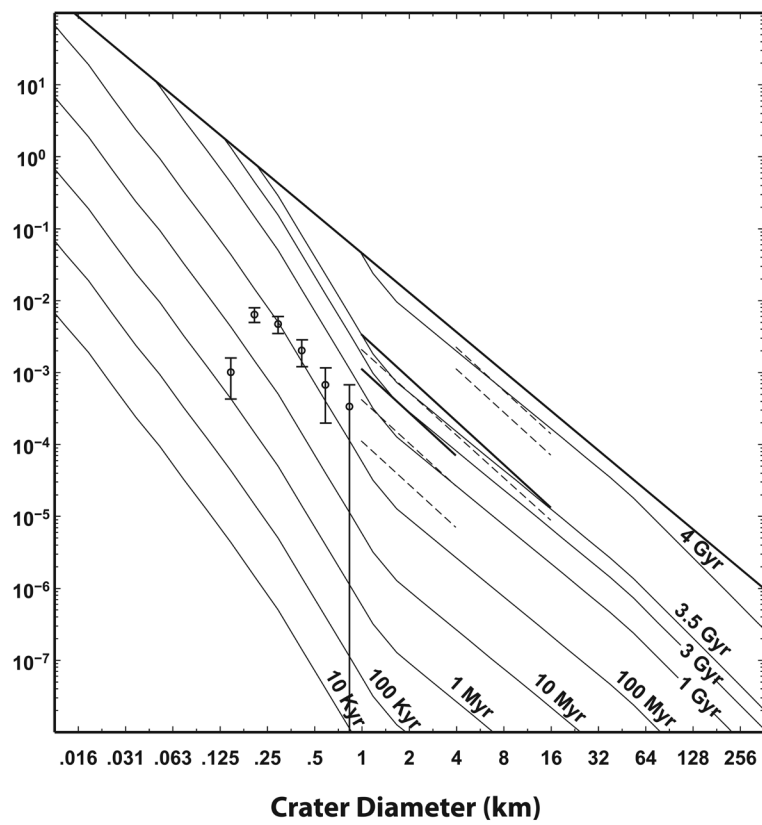
Chapter 2, Figure. 11



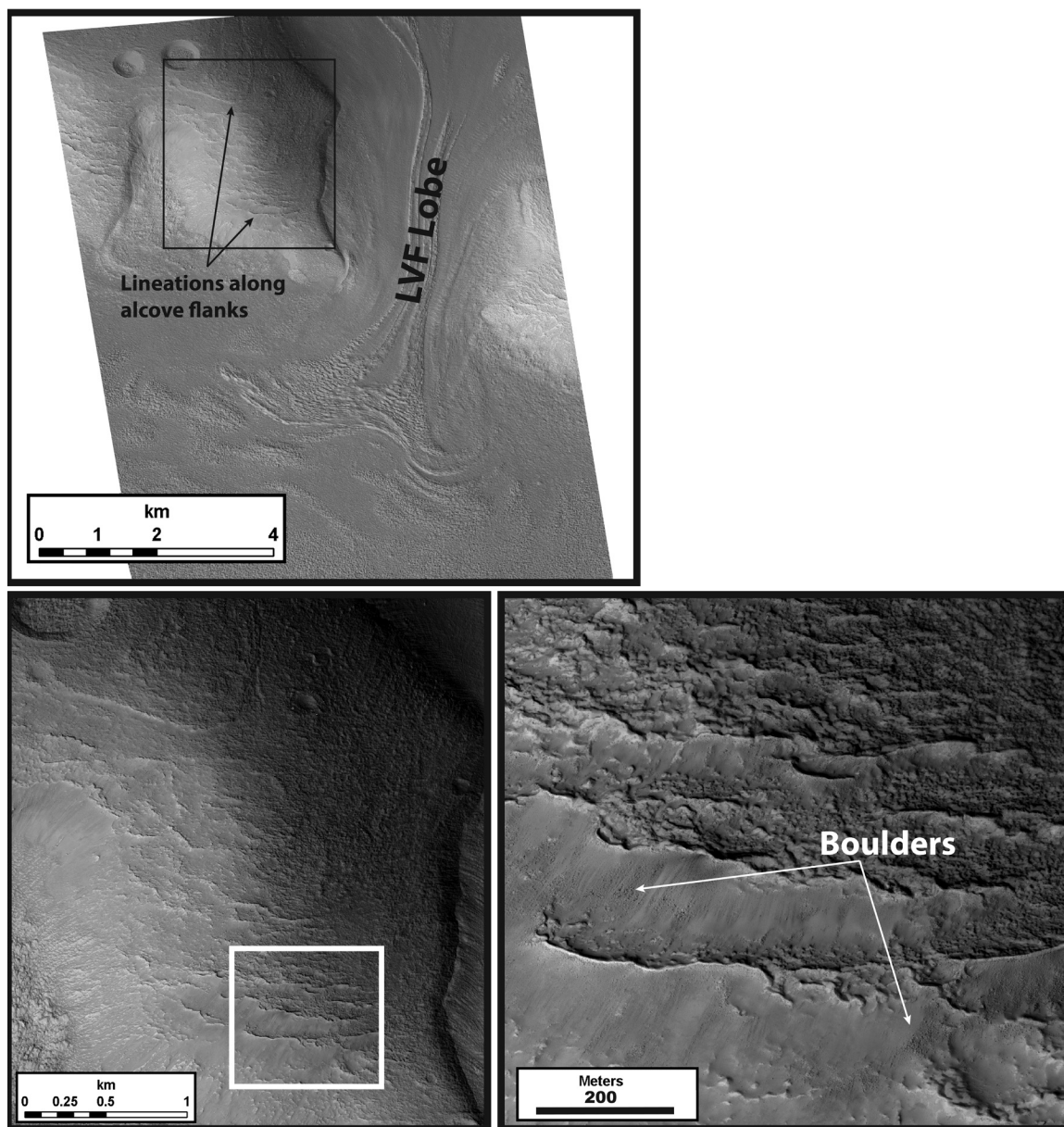
Chapter 2, Figure. 12



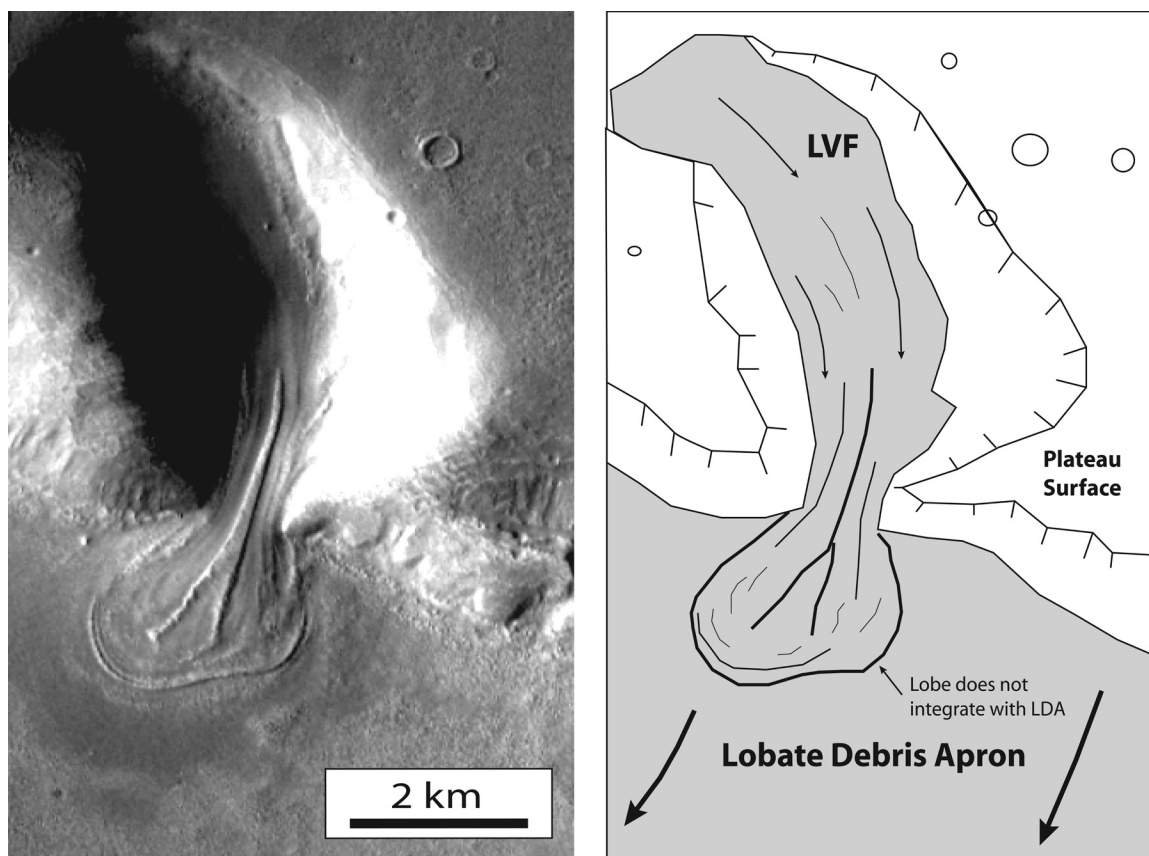
Chapter 2, Figure. 13



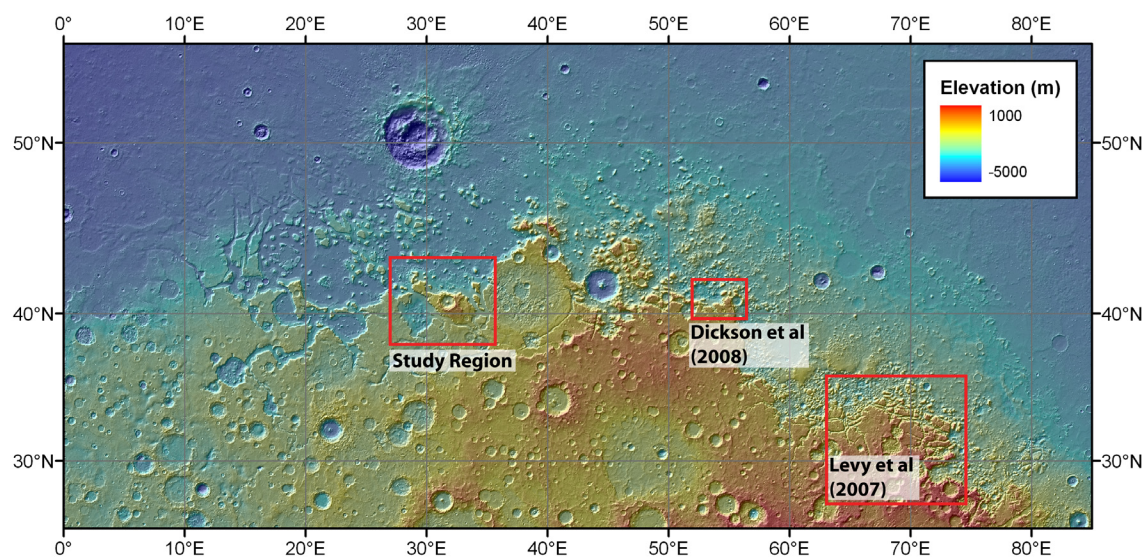
Chapter 2, Figure. 14



Chapter 2, Figure. 15



Chapter 2, Figure. 16



Chapter 2, Figure. 17

Chapter Three

Formation of Gully Components in Wright Valley, Antarctic Dry Valleys: Implications for Mars.

Gareth A. Morgan

Abstract

The McMurdo Dry Valleys of Antarctica are a hyperarid, cold polar desert in which precipitation is limited to minimal snowfall that is generally exceeded by sublimation. Despite the frigid conditions, annual streams, supplied by meltwater from cold-based glaciers, do flow for limited periods each summer and drain into open and closed basin lakes. In addition to the streams, smaller scale, spatially-confined gully systems are also observed. These experience short-term fluvial activity, but have not attracted the same degree of scientific attention and thus their relative contribution to the hydrological and ecological systems of the Dry Valleys remains to be fully established. These gully systems are also morphologically very similar to young erosional features recently discovered on Mars, and thus provide a natural laboratory in which to test martian gully formation models. Here we report on the detailed investigation of two gully systems in Upper Wright Valley. This microenvironment marks the driest and coldest region of meltwater formation in the Dry Valleys and thus is the most similar to martian conditions during the Amazonian. Our results demonstrate that unlike the stream systems, gully activity occurs due to the solar insolation-induced melting of both perennial and annually deposited snow deposits. This processes occurs during daily periods of peak insolation. Strong winds redistribute the low amounts of snowfall during the winter and preferentially deposit it in topographic lows. The gully channels represent effective wind traps that concentrate snow into banks tens of centimeters thick. The melting of this snow was observed to generate a range of fine-scale fluvial features. Similar features have been identified in sub-meter resolution images of martian gullies. Thus we propose that gully activity in the Dry Valleys provides a compelling model for the formation of several aspects of martian gullies.

1. Introduction

The McMurdo Dry Valleys of Antarctica represent an ice-free region of the Transantarctic Mountains, situated between the East Antarctic Ice Sheet and the Ross Sea. The Dry Valleys are a hyperarid, cold polar desert (Marchant and Head, 2007), characterized by mean summer temperatures below freezing and winter lows of $< -40^{\circ}\text{C}$. Precipitation throughout the year is restricted to snowfall ($< 10 \text{ cm yr}^{-1}$, Bromley, 1985), which is exceeded by sublimation, and as a consequence severely limits the availability of liquid water. Despite the frigid conditions, small streams and gully systems are active for limited periods during the Austral summer (Conovitz et al., 1998). The streams are sourced by meltwater from cold-based glaciers, which melt during periods of direct solar radiation incident on the glacier faces (Fountain et al., 1998). In many instances the streams feed perennially ice-covered lakes that represent the base level of closed hydrological systems (Conovitz et al., 1998). Gully systems are present along the steepest slopes of the valley walls and are distinguished in this study from the streams by their limited channel length, which terminate as depositional fans rather than extending for a significant length along the low slopes of the valley floors, as is the case with the streams.

The gully systems present in the Dry Valleys are morphologically very similar to youthful gully features observed on Mars in high-resolution images (Malin and Edgett, 2000) (Fig 1). Gullies on Mars are defined by a tripartite morphological classification system consisting of an alcove, channel(s) and depositional fan (Malin and Edgett, 2000). The Dry Valleys gullies also exhibit these units (Fig.1) and thus this classification system can be applied as a basis for the identification and cataloging of gully systems in Antarctica.

The discovery of gullies on Mars has been a source of immense scientific interest because their presence suggests liquid water has flowed on the surface in the recent geologic past, under conditions previously considered incompatible with melting and surface runoff due to conditions below the triple point of water (e.g. Carr, 1996). Although theories have been put forward to

explain the formation of the gullies in the absence of water (i.e. dry flows: Treiman, 2003; liquid CO₂ flows: Musselwhite, 2001), the majority of the community are in agreement that the latitudinal dependence of gully distribution (poleward of $\sim 30^\circ$ in both hemispheres, Malin and Edgett, 2000; Heldmann and Mellon, 2004; Costard et al., 2002) and the occurrence of sinuous channels (Malin and Edgett, 2000) with internal fine-scale fluvial features (McEwan et al., 2007) is most consistent with erosion by liquid water. Many questions still remain, and include: 1) What is the origin of the water that carved gullies? 2) What is the mechanism of erosion that has formed the gullies (i.e. what is the water/rock ratio during formation)? 3) What is the timing and duration of gully activity? Dating by Reiss et al (2004) and Schon et al (2009) has shown recent gully activity to have occurred ~ 1 Ma, and morphological evidence suggests that multiple episodes of emplacement are involved in gully formation (Schon et al., 2009). Malin et al (2006) have also shown evidence for activity in gullies in the last decade. This raises the questions: how long then have gullies been forming, and are they likely to be active again?

In terms of precipitation, mean annual temperature and geological setting, the Dry Valleys have long been considered the most Mars-like of terrestrial environments (e.g. Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005). Therefore, research in Antarctica offers an important natural laboratory in which to monitor gully activity in real time and test models of martian gully formation. Additionally, the ~ 0.3 m/pixel spatial resolution of the High Resolution Science Experiment (HiRISE) camera currently in orbit around Mars has made it possible to make direct comparisons between field observations and martian surface features. Due to the extreme aridity of the Dry Valleys, the presence of liquid water, even in a transient state is of important scientific consequence to terrestrial research. Perennial ecosystems consisting of microbial communities that can survive long periods of desiccation can support relatively high biomass in the presence of seasonal sources of water in the Dry Valleys (McKnight and Tate, 1997; McKnight et al., 1999). The relative stability of water can also be used to define local microclimates and their migration can be used as a proxy for climatic change

(Marchant and Head, 2007). Much of the work to date has focused on the stream systems associated with glacial melt (Chinn, 1993, Fountain et al., 1998, Conovitz et al., 1998; McKnight et al., 1999) with less attention devoted to defining the relative contribution of gullies within the Dry Valley system. Here we report on the results of a field campaign conducted during the austral summer of 2006-7 in the Dry Valleys to investigate and document the initiation, duration and nature of gully activity.

2. Field Study Site and the Anatomy of Antarctic Gully Systems

Within the Dry Valleys, moisture and temperatures both decrease as a function of distance inland from the Ross Sea and with the corresponding increase in elevation, producing a pronounced climate gradient. Marchant and Head (2007) subdivided the Dry Valleys into three microenvironments, each of which has distinctive geomorphic characteristics that reflect this gradient. The majority of the Dry Valleys surface consists of unconsolidated sediment (e.g., colluvium, till) modified by contraction-crack polygons. Ice-cemented permafrost occurs in most places throughout the Dry Valleys and is most commonly encountered at depths of 0-70 cm. Above the ice table, a wet active layer is seen in the warmer coastal microenvironment zone. Summer subsurface soil temperatures exceeding 0°C permit thaw and active layer formation, and the zone receives the highest snowfall of the Dry Valleys. A dry "active" layer occurs in the intermediate zone, although the zone does experience localized meltwater production. In the stable upland zone, soil temperatures generally fail to rise above 0°C and the zone experiences negligible quantities of snowmelt and thus lacks active layer formation.

In order to best constrain the martian conditions we concentrated our efforts within the South Fork region of upper Wright Valley (Fig. 2). This represents the most elevated (and hence driest) portion of the intermediate microclimate zone, where it borders the colder and more Mars-like stable upland zone. South Fork consists of a 2 km wide valley formed between the northern

slopes of the Asgard Range and the Dias. Local relief is > 1 km. This region marks the most inland extent of active fluvial features within the Dry Valleys and is most analogous to Mars during the Amazonian in terms of precipitation and maximum surface temperature values. Unlike the Dry Valley streams, there are no significant adjacent glacial systems to supply meltwater to the South Fork gullies, and so the most significant ice deposits within the study region are snowbanks, as established by long-term photographic records. Meteorological stations established as part of our fieldwork recorded diurnal air temperatures that oscillated in response to solar radiation around 0°C , (Fig. 3).

Our work focused on two active gully systems situated along the southern wall of the Asgard range to the East of Don Juan pond (Fig. 2c, 4). The larger of the two gullies, referred to as the ‘east gully’ originates in an alcove cut into the Asgard Range, ~ 1 km above the valley floor (shown in Fig. 2). The gully channel is eroded into a layer of colluvium and extends downslope for over 2 km before terminating in fan deposits. The channel is $\sim 2 - 3$ meters wide and displays a prominent ‘v’-shaped cross section and levees for the majority of its length. Closer to the fan on the shallower slopes (Fig. 4) the channel floor opens up and becomes ~ 2 meters wide. The upper portions of the channel consist of angular boulders $\sim 0.30 - 1$ m wide, but the channel floor becomes progressively filled with fine-grained sediments as a function of distance downslope (Fig. 2, 4).

The upper portion of the second gully system (fig. 2), refereed to as the ‘west gully’ is located along the surface of a dolerite tongue that originates within an alcove adjacent to the east gully alcove (Fig. 2, 4). Multiple tributaries that begin within the tongue feed the main channel of the gully, which is ~ 2 m across at its widest section. Both gully channels show clear evidence of interaction with the troughs of sand wedge polygons that cover the valley walls (Levy et al., 2008; 2009) (Fig. 4).

Two distinct deposits of snow were identified as the source of meltwater and activity within the two main gully systems studied in South Fork. The first was derived from annual

snowbank deposits that had accumulated in the gully channels. The second was from large-scale (>100 m) perennial snow banks within the gully alcoves. Evidence for groundwater discharge has been reported in the Dry Valleys as a potential input for the lakes (Doran et al., 2007) and the Dry Valleys drilling project also reported the presence of a liquid water reservoir below the permafrost in proximity to Don Juan Pond in South Fork (McGinnis, 1981). With regard to Mars this is an important consideration as much discussion within the literature has been devoted to the source of gully activity, which has divided the community into two camps. The original model of gully formation (Malin and Edgett, 2000) proposed that perched aquifers were the source of gully activity; a groundwater source negates the need for conditions conducive to melting and formation of stable runoff on the martian surface. The alternative view is that gullies are supplied by water from the melting of atmospherically derived snow/ice deposits (Costard et al., 2002; Christensen, 2003; Dickson et al., 2007; Head et al., 2008). Hence, one of our objectives was to search for evidence of groundwater sapping as a potential source of gully activity in South Fork. Despite extensive searches no spring features were identified during our field campaign and the correlation between snowmelt and gully activity (which is discussed in more detail below) rules out a groundwater source as a significant contributor to the present-day gully systems.

In regional soil mapping of South Fork by McLeod et al (2007), the presence of ice cemented permafrost at shallow depths (< 70 cm) across the southern valley wall, is attributed to meltwater derived from high elevation. In our more detailed local analysis, we found the depth of the ice table in proximity to the gullies to be ~20 cm. Because this ice source is so close below the gullies, specific consideration was given to establishing if the melting of the uppermost ice layer was a potential source of water for gully erosion. To investigate this, temperature probes were installed along the ground surface and at progressive depths of 5 cm increments within the active layer to the surface of the ice table (Fig. 5). This demonstrates the effect of solar heating on the surface and subsurface over the time period of observed gully fluvial activity. Although diurnal temperature variations were experienced along the ice table surface, the zero isotherm did

not propagate to this depth, preventing the occurrence of melting, and indicating that melting of ground ice was not an important source in the observed gully activity. Our data does not rule out ground-ice thaw as a contributor to meltwater and gully systems fluvial activity during warmer years, such as occurred in the summer of 2001 – 2002 (McLeod et al., 1997).

Below we report on the two main sources of snow melt that were directly observed to supply surface water runoff and erosion within the two gully systems studied in South Fork, and we evaluate their relative contributions.

3. Annual Snowbank Deposits within the Gully Channels.

Despite the low precipitation levels, the snow that does fall within the Dry Valleys and does not sublimate is redistributed by the strong winds and preferentially deposited and trapped in topographical hollows. This includes the gully channels themselves, which act as efficient snow traps (Fig 6) demonstrating a potential positive feedback effect in gully erosion. At the beginning of the season (during late November) the lower gully channels were almost completely filled with snow deposits, tens of centimeters deep. At this time the west gully had an estimated 320 m³ of snow within its 400 m long main channel, a value estimated from an average snowbank depth of ~40 cm and a width of ~2 m. This estimate does not include snow accumulation within the multiple tributaries and interconnected polygon troughs that feed the channel. Assuming a snow density of ~ 400 kgm⁻³ which corresponds to wind packed snow, this volume of accumulated snow provides a potential ~ 130 m³ (1 x 10⁵ litres) of meltwater to the system.

We observed that the generation of meltwater was strongly controlled by the intensity of solar radiation. Despite 24 hour sunlight during the Austral summer, the sun does descend below the >1 km high valley walls for ~ 8 hours during the day (Fig 7). During this period the illumination of the gullies is limited to scattered light. This period corresponds to the diurnal periods of minimum solar radiation (~15 Wm⁻²) in Figure 7. When the sun rises above the valley

walls, the insolation rapidly increases and conforms to a sinusoidal pattern corresponding to the zenith angle of the sun's position in the sky (Fig. 7). During periods of peak insolation, temperatures rose sufficiently for melting of the snowbanks to occur (Fig. 8a). This period of melting typically lasted for ~ 7 hours (Fig. 7). The meltwater was initially observed to infiltrate into the dry active layer above the ice table. When the active layer became locally saturated or infiltration capacity was exceeded, surface runoff was initiated within the gully channel (Fig. 8a). When the air temperature dropped below zero degrees Celsius, the melting ceased and the surface runoff typically froze in situ (Fig. 8b).

The formation of runoff was accompanied by the development of a hyporheic zone that caused a darkening of the soil (due to moisture) that extended both laterally and directly ahead of the flow front of surface runoff (Fig. 9). The hyporheic zone can be described as the immediate area underneath and adjacent to surface runoff, in which there is a downslope movement of water in the active layer that is in direct communication with the water in the main channel (McKnight et al., 1999). This permits the bidirectional flux of water and solutes between the active layer and the channel flow. The spatial extent of the hyporheic zone as it was observed from the surface due to the associated decrease in the albedo was significant (> 2 m in both directions from the channel in the lower sections of the gully, Fig. 8c, 9). It has been noted that for the larger stream systems sourced by glaciers in the Dry Valleys (e.g. Conovitz et al., 1998) the hyporheic zone provides a reservoir for a substantial volume of the generated meltwater. The infiltration and storage of water in the active layer can also facilitate geomorphic work. Small-scale lobe forms were observed at the distal margins of the gully channels and are potentially related to solifluction initiated by the formation of the hyporheic zone. Features of this scale, if sufficiently preserved, could be identifiable within martian gullies at HiRISE resolution, and should be searched for in future surveys.

The sensitivity of meltwater production and gully erosion to insolation was further demonstrated by the effect of clouds. The occurrence of clouds can be seen directly in the solar

radiation data (Fig 7) as sharp troughs in the insolation values. For example, compare the 5th of December (Fig 7), a cloud free day, with the 4th, a day which experienced a cloudy afternoon. The high frequency of cloud cover or overcast conditions prevented the generation of meltwater. This is demonstrated in Figure 10, which shows channel activity in the west gully on two separate days around 1 pm local time. Under overcast conditions on the 1st of December, the resulting lack of high solar radiation prevented air temperatures from rising sufficiently to cause melting, and the channel was observed to contain ice from the previous days activity (Fig. 10a). In contrast, the 2nd of December experienced relatively fewer clouds and as a result enhanced melting and surface runoff flow occurred (Fig 10b). The larger Dry Valley streams that are also dependent on insolation for meltwater production (from glaciers) mirror this variation in gully activity, and variations in streamflow of an order of magnitude have been recorded during a single days flow (Conovitz et al, 1998; McKnight et al., 1999). In Figure 7, note the occurrence of spikes in solar radiation (e.g. on the 4th in Fig. 7) that are above the maximum values encountered during cloud free days (i.e. the 5th in Fig. 7). This can also be the result of clouds. The edges of scattered cumulus clouds in proximity to the sun in otherwise clear-sky conditions can reflect additional solar radiation to the ground, and this mechanism is particularly effective in dry conditions (Gates, 1980). The potential contribution of this effect to the generation of snowmelt is not fully known, although it is a worthwhile consideration for future work.

The runoff generated by the meltwater from the snowpacks carved multiple adjacent small braided channels ~ 1 cm across that occupied active streams 0.5 – 1.5 meters wide (Fig. 8a,c, 10b). Peak velocity was estimated at 0.1 ms⁻¹ corresponding to discharge of ~ 1x10⁻³ m³s⁻¹. This level of flow was only capable of transporting sand-sized particles and the water had a cloudy appearance suggesting the additional transportation of material in suspension and solution. Despite the low flow, terraces were cut in the floor of the gully channels (Fig. 11). Over the course of the gully activity period, flights consisting of up to four or five terraces (Fig. 11b) were present within the portions of the gully channels that experienced the highest levels of erosion. In

these locations the net incision of the channel floor was ~ 0.3 m over the observed time period (Fig. 11a). The formation of the terraces is the result of discharge variations as a function of diurnal variations in insolation (Fig. 7). The initiation of melting each day and the subsequent daily increase in discharge to periods of peak daily insolation caused an associated increase in erosional energy of the stream flow. This prevented the establishment of equilibrium conditions and instead permitted the down cutting of the gully channel floor. This was particularly effective on days with clear skies that followed cloudy conditions (e.g. the 5th of December, Fig. 7). The terraces displayed fine-scaled horizontal stratification (~ 5 mm thick layering) in cross-section, consisting of sand-grain sized material (Fig. 11b). These layers likely represent alluvium deposited by stream flow, potentially during the close of activity at the end of the previous day, when declining melt production switches the dynamics of stream flow from a degradational to an aggradational regime. A substantial component may also originate from aeolian sediments deposited by the strong winds during the winter, in a manner similar to the deposition of snowbanks. The incision of the gully channel revealed buried deposits of snow and ice that also contributed to melt production upon exposure to the surface (Fig. 11). These deposits represent wind blown snow banks that had not been fully melted during previous summers and had subsequently been buried by aeolian sediments, further demonstrating the contribution of the wind and wind-blown snow in gully activity.

Full ablation of the snowbanks within the lower sections of the gully channels took approximately two weeks. Surface runoff ceased to occur prior to the end of this two-week period, as the snow lost an affective ‘critical mass’ for runoff generation and the remaining melt only contributed to infiltration into the active layer. Repeated observations over the active meltwater generation period permitted a study of the nature of snowmelt. The dynamics of snowmelt provides the core of several martian gully formation models (e.g. Christensen, 2003; Williams et al., 2009) and thus monitoring its occurrence in real time is essential for constraining future snowmelt models and assessing their application to meltwater conditions. The initial stage

of ablation was in the form of long (tens of meters), 10 cm wide sinuous channels that were carved into the surface of the snowbanks themselves (Fig. 6b). These features are similar to the ‘dirty snow gullies’ reported on Devon Island by Lee et al (2006), with the exception that the South Fork examples are several times the length of those on Devon Island. The presence of singular channels that extend over distances > 10 m suggest that they were formed by meltwater on the surface of the snowbanks. This activity was never directly observed, occurring prior to our arrival on November 28th. The small scale of the channels implies that only small amounts of melt water would be required to carve them by thermal erosion. The initial main phase of ablation was concentrated along the top of the snowbanks, and resulted in the generation of meltwater.

Excavations into snowbanks revealed layers of ice that formed horizons at various levels within the snowbank. This suggests that melting occurred at and/or close to the surface of the snowbank and meltwater percolated down through the snow prior to refreezing at depth within the snowbank. Hence a component of the water that generates the initial gully flow originated from the melting of the upper portions of the snowbanks. The ablation of the underside of the snowbanks was intensified by initiation of surface runoff, which was especially effective when the meltwater of an upstream snowbank made contact with the base of separated snow deposit downstream. The prolonged generation of meltwater effectively hollowed out the interior of snowbanks. The ablation of the surface of the snow deposits led to the formation of 10 cm diameter ‘sun cups’ (Fig. 6). This was typically accompanied by the formation of a lag of fine-grained dark sandy material on the surface of the snow (Fig. 6). The origin of this lag layer was determined to be from the combination of dust originally mixed in the snow being concentrated at the surface through sublimation (sublimation lag) and direct deposition of wind-blown fines that became trapped by the surface topography of the snow. The accumulation of this dust and sand on the surface of the snow reduced its albedo and further assisted the melting processes. This is potentially an important effect with regards to martian gully formation. Due to the thin atmosphere on Mars (surface pressure currently one hundredth that of Earth), the air temperature

remains extremely cold and thus surface temperatures are controlled by the albedo and thermal properties of the surface (Hecht, 2002). Hence a mechanism for reducing the albedo of martian snow banks provides potential insight into the generation of meltwater on Mars. Modeling work by Williams et al (2009) of a martian snowbank under enhanced obliquity conditions has demonstrated that the generation of a sublimation lag consisting of dust is required for melting to occur.

In addition to fluvial erosion along the gully channel floors, the steep slopes of the channel walls made them vulnerable to erosion from mass wasting (Fig. 12). This was observed to occur along the steeper portions of the southern wall of South Fork (slopes $>20^\circ$), where the gullies exhibit v-shaped cross sections and are > 1 meter deep (Fig. 12). The gully walls consist of colluvium containing a mixture of material ranging from sand-sized particles to small boulders ~ 30 cm in diameter (Fig. 12). Cracks, several centimeters wide were present parallel to the gully channels along the surface of the slopes. These cracks served to isolate blocks of colluvium ~ 1 m long by 0.5 m wide, which were prone to collapse into the gully channel (Fig. 12a). The cracks appeared to be linked to the troughs of thermal contraction crack polygons that cover the northern slopes of South Fork (Fig. 4). The opening of polygon troughs adjacent to the gully channel due to thermal contraction during the summer will provide a force on the channel wall material in a direction of the channel itself. Therefore, polygon formation may assist the mass wasting of the gully channels (see Levy et al., 2008 for more details on polygon-gully interactions).

This provides a mechanism for the lateral expansion of gully channels in the absence of fluvial activity, and would be expected to occur regardless of the amount of snowfall that was available for meltwater generation. However, in order for the v-shaped cross section of the channels to be maintained at the angle of repose (Fig. 12b), erosional processes must have been in operation along the floor of the gully channel and must have kept pace with mass wasting through the downslope movement of material. In some instances material deposited in the channels from mass wasting caused the burial of snowbanks, and provides another mechanism for the formation

of buried ice within the channel floors. The fines from within the fallen blocks provide a potential source of fine-grained material that could be transported by active stream flow and deposited within the lower sections of the gully channels systems, contributing to the formation of the depositional fans. The preferential removal of fine-grained material from the higher and steeper ($>20^\circ$) portions of the gully channels and its subsequent deposition close to the valley floor, where slopes are significantly lower ($\sim 5^\circ$) would explain the higher concentration of boulders and rocks in the elevated portion of the gully channels relative to the lower channels.

4. Perennial Snowbanks within the Alcove.

Analysis of multi-year sets of Dry Valley satellite images and aerial photographs allowed us to assess the spatial duration and longevity of snowbanks within South Fork. This revealed a concentration of perennial snowpacks, tens of meters wide within the alcoves and alcove-like terraces and depressions along the margins of the northern edge of the Asgard Range that forms the southern wall of upper Wright Valley (Fig. 2). The alcoves are cut into doleritic sills that cap the summit of the slopes. The elevation of the alcoves (~ 1 km above the floor of South Fork) creates differential meteorological conditions relative to the valley floor (and lower portions of the gully channels). As a result of the increase in altitude, the associated adiabatic expansion of moist air originating down valley from the Ross Sea can be sufficient to enable cloud formation and precipitation (Fig. 13). During storm events low-lying clouds were observed to hug the alcoves and preferentially deposit snow within them (Fig. 13c). The snowfall during these events was minimal (~ 1 mm), but the accumulated effect of the precipitation within confined alcoves (hence protected from the wind) over multiple year periods can account for the building up of large-scale snow deposits. The eastern gully system originates from a 750 m wide alcove that contains a patch of snow and underlying ice ~ 500 m wide (Fig. 14a). A small ~ 50 cm wide channel emanates along the edge of the snowbank and connects with the main gully channel

at the steepest portion of the slope at the apex of the alcove (Fig. 14b, c). The alcove walls and floor were draped in a weathered accumulation of angular fragments of sandstone ~20 cm long. The small channel was incised into this coarse regolith layer, though the bed of the channel did contain finer grained material consisting of sand-sized particles.

By early December 2006, marginal parts of the snow bank had started to melt, and the adjacent channel was occupied by water actively flowing from the ice toward the scarp and cascading over the cliff. As was observed in lower portions of the gully channel, the small channel was ice covered in the mornings and began to melt under direct solar insolation during the day. During peak flow the water was observed to be cloudy, but the sediment load was minimal and largely confined to bed load movement of sand-sized particles. Subsequent to the flow of water over the scarp, the water was lost within the coarse bouldery deposits (D: ~ 50 cm) on the steepest and most inaccessible parts of the cliff. End to end flow from the alcove down the main channel to the fan was not directly observed during the 2006-7 field season, although it would be difficult to observe due to the large boulders that fill the upper half of the main channel. Alternatively temperatures in the alcove may simply not have been maintained high enough for sufficient durations to generate the required volume of melt from the perennial snowbanks to permit continuous, end-to-end channel flow.

5. Discussion and Application to Mars

1. Sources of Meltwater in Gully Formation: Gully activity in the Antarctic Dry Valleys results from a combination of the melting of perennial and annual deposits of snow. The annual snowbanks were not formed in situ (by direct precipitation) but originated from the redistribution and concentration of snow within the gully channels by the action of the wind. This demonstrates that complete snow cover is not required to form gullies; wind-blown snow in a region of extremely low annual precipitation is sufficient to form significant local snowpack

accumulations. Martian GCM (e.g. Mischna et al., 2003; Madeleine et al., 2007) have demonstrated that snow is likely to form on Mars during periods of higher obliquity as a result of the redistribution of water ice from the polar caps to mid latitudes.

2. Timing and Duration of Melting: Snowmelt (and hence gully activity) only occurred during periods of peak daily insolation, when daily temperatures were at their maximum, and thus only lasted for ~7 hours per day. The duration of days in which surface runoff was observed was less than 2 weeks. Therefore, gullies in Antarctica are only active for 0.005% of the entire year. Hence, models of martian gully formation should consider peak climatic conditions rather than annual values of temperature and pressure.

3. Role and Importance of Hyporheic Zones: Gully activity was accompanied by the formation of hyporheic zones that represent both a significant sink and storage for meltwater in an otherwise desert environment. Considering the desiccated nature of the upper subsurface layer martian regolith at gully forming latitudes, any runoff generated is likely to infiltrate into the ground (depending on ground temperatures) unless there is a mechanism to prevent it. The exchange of salts between the subsurface and the active channel flow may provide a mechanism for permitting freezing point depression within surface runoff, which could enable longer duration of flow. This may be of significance on Mars in light of the recent discovery of perchlorates in martian regolith (e.g. Hecht et al., 2009).

4. Application to gullies on Mars: The occurrence of small-scale fluvial erosion by snowmelt and the generation of braiding, terraces and streamlined features is morphologically similar to internal channel features found within the main channels of some martian gullies (McEwan et al., 2007, Fig. 15). This suggests that some of the gullies on Mars may have formed by active stream flow comparable to that generated by snowmelt in Antarctica. The mass wasting of the main channel walls demonstrates how gullies can develop through a range of erosional processes and thus may have complex formation histories.

As the snowbanks ablated, a surface lag of dust formed on their surface through a

combination of the sublimation of dusty snow and the accumulation of dust blown directly on to the snow (Fig. 6). This reduced the albedo and assisted further melting. This process has been suggested to occur on Mars as a result of thermal modeling of snowbanks at high obliquity (Williams et al., 2009).

Gully activity in the Dry Valleys demonstrates the importance of snowmelt as a source of water in a hyperarid, cold polar desert environment, and has significant consequences for the local hydrological and ecological systems. We argue that the generation of surface runoff through the melting of surficial deposits of snowbanks within gully systems in the Antarctic Dry Valleys provides a compelling model for gully activity on Mars. Further research is required to assess the full spatial distribution of currently active gullies and address the inter-annual variation of activity, so that their full hydrological contribution to the Dry Valleys relative to the stream systems can be constrained.

Acknowledgements

Special thanks are extended to my advisor **James W. Head** and to **David R. Marchant**, **James L. Dickson** and **Joseph S. Levy** for extensive discussions on this work during the formulation of the ideas and content of this chapter, and for their support with the fieldwork conducted in Antarctica. I look forward to additional assistance and collaboration with all of them as this contribution is prepared for publication.

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Figure Captions

Fig. 1. Comparison between a gully within Wright Valley in the Antarctic Dry Valleys (left) and gully systems on Mars (Right). On both planets the gullies consist of alcoves, sinuous channels and deposition fans, which form the basis of a gully classification system for Mars

(Malin and Edgett, 2000). Antarctic image was taken from helicopter during 2006 field season. Mars image, subset of HiRISE image: PSP_001842_1395.

Fig. 2. Context images of the Study Site. A. South Fork (highlighted by white box) in the context of Wright valley. B. South Fork of upper Wright Valley. The triangle shows the perspective of the photograph in C. C. Field Photograph of the two gully systems (labeled East and West) investigated. Note the occurrence of snow in the alcoves and channels. The horizontal white line in the lower right corner highlights our field camp for scale. The elevation of the summit of the Asgard Range is 1000 m above the floor of South Fork.

Fig. 3. Solar radiation and air temperature measured by a meteorological station positioned on the lower portion of the west gully channel for the austral summer of 2006 - 7. The black bar represents the data in Fig. 7, which was the period of maximum gully activity.

Fig. 4. View of the lower channels and fans of the two gullies studied. The image was taken on the 29th of December, 2006, after much of the snow had melted within the channels. Progradation of the fan of the western gully by the main gully channel is evident. This is interpreted the result of the positive erosional feedback effect of the gully channel trapping snow which aids the further erosion of the channel and fan deposit. The sandwedge thermal contraction crack polygons that cover the slopes adjacent to the gullies can be clearly seen, and appear to have influenced gully channel development through the creation of troughs that can concentrate drainage and flow (Levy et al., 2008; 2009). Arrow points to North.

Fig. 5. Surface and subsurface temperatures recorded close to the lower portion of the west gully channel during the period of maximum gully activity. The ice table at this location was at a depth of 20 cm. The plot shows that the zero degree isotherm did not propagate down to the

ice table surface over this time period. Hence, the melting of subsurface ice did contribute to gully activity.

Fig. 6. Gully channels completely filled with wind blown snow in late November, early December. Note the lack of snow on the slope surfaces adjacent to the channel. The action of the wind provides a mechanism for concentrating the very low amounts of annual snowfall. A. Large patches of dirty snow and suncups can be seen on the surface of the snowbank. This originated from the formation of dusty lag deposit. B. Small sinuous channels can be seen on the surface of the snow bank.

Fig. 7. Solar radiation and air temperature measured by a meteorological station positioned on the lower portion of the west gully channel over the period of maximum gully activity. Note, the solar radiation data for the 5th of December represents cloud free conditions. Deviations from this pattern on other days are caused by clouds. The green bar represents the time the photograph in Fig. 10a was taken, the purple bar represents the time the photograph in Fig. 10b was taken.

Fig. 8. Snowmelt, surface runoff and erosion. A. Melt from the snowbank on the right hand side of the photograph is actively forming the fine-scale fluvial features. B. Frozen surface runoff that forms during periods of low solar radiation. C. The fluvial depositional and erosional features from snow meltwater flow. Note that despite the lack of flow when the photograph was taken, the hyporheic zone is still visible as a dark region around the channel.

Fig. 9. Development of surface runoff and hyporheic zone. A – D represents a time sequence spanning ~ 30 mins. The surface runoff followed the initial lead of the hyporheic zone, which can clearly be seen in the most distal lobe of flow. The reason for this is that the dry active

layer had to become locally saturated or the infiltration capacity exceeded before runoff could occur.

Fig. 10. Photographs taken around 1pm of the western gully on subsequent days (highlighted on the solar radiation plot in Fig 7). A. 1st of December 2006: Overcast conditions (Fig. 7) prevented the re-melting of runoff frozen from the previous days flow. B. 2nd of December 2006: Lower cloud levels permit melt generation and surface flow.

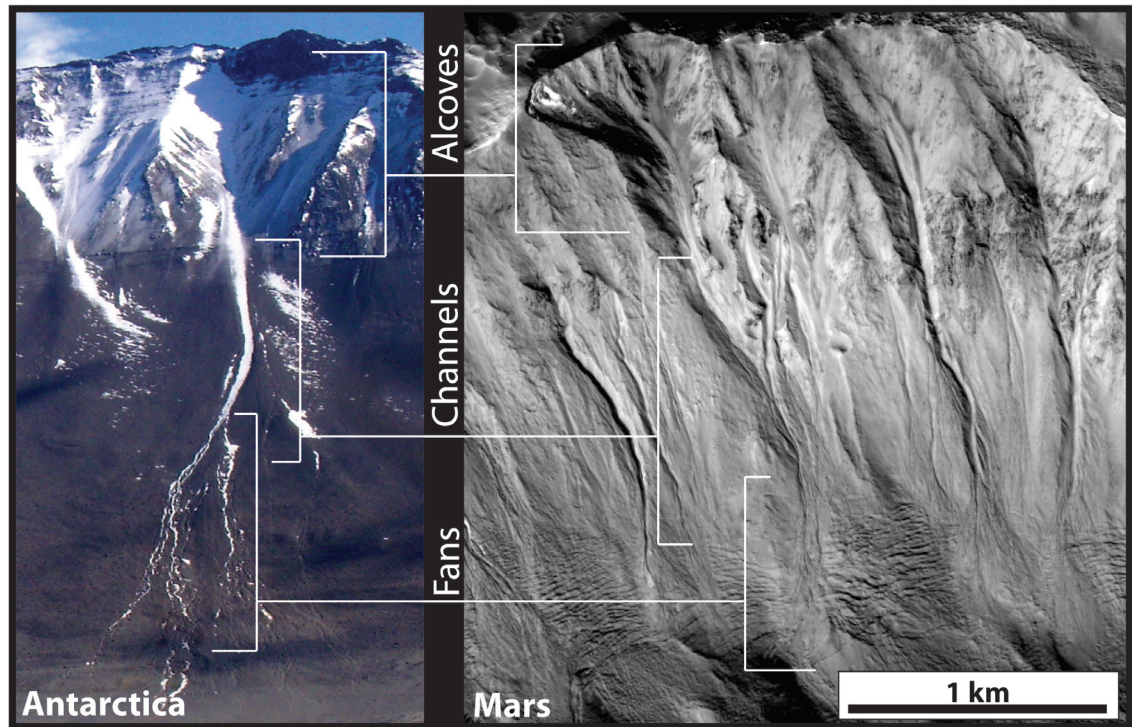
Fig. 11. Terraces developed from progressive incision due to fluvial erosion. Variations in melt production, will occur both on a daily basis (due to the sun's trajectory in the sky and the influence that this has on insolation for a fixed position on the surface, and due to the effect of clouds) and also on a day to day basis (due to the effect of clouds). This will influence discharge and change flow from an aggregational to a degradational state, permitting incision. A. Previous gully channel floor level (note small scale erosional features similar to Fig. 8) is left stranded as a terrace > 20 cm above the current level of flow to the right of the image. Also note buried ice is being exhumed by flow on the left of the image. B. Cross section perspective of a flight of terraces. Note fine lineations of sand-sized particles.

Fig. 12. Mass wasting along upper portions of the east gully channel. A. 50 cm block of colluvium is ready to fall into the gully channel. The crack separating the block from the slope is common along the gully channel and may be related to thermal contraction crack polygon margins. Note heavily ablated snowpack and associated meltwater-induced runoff. B. View looking up the channel of the east gully. Note the 'v' shaped cross section of the gully channel, which is prone to failure.

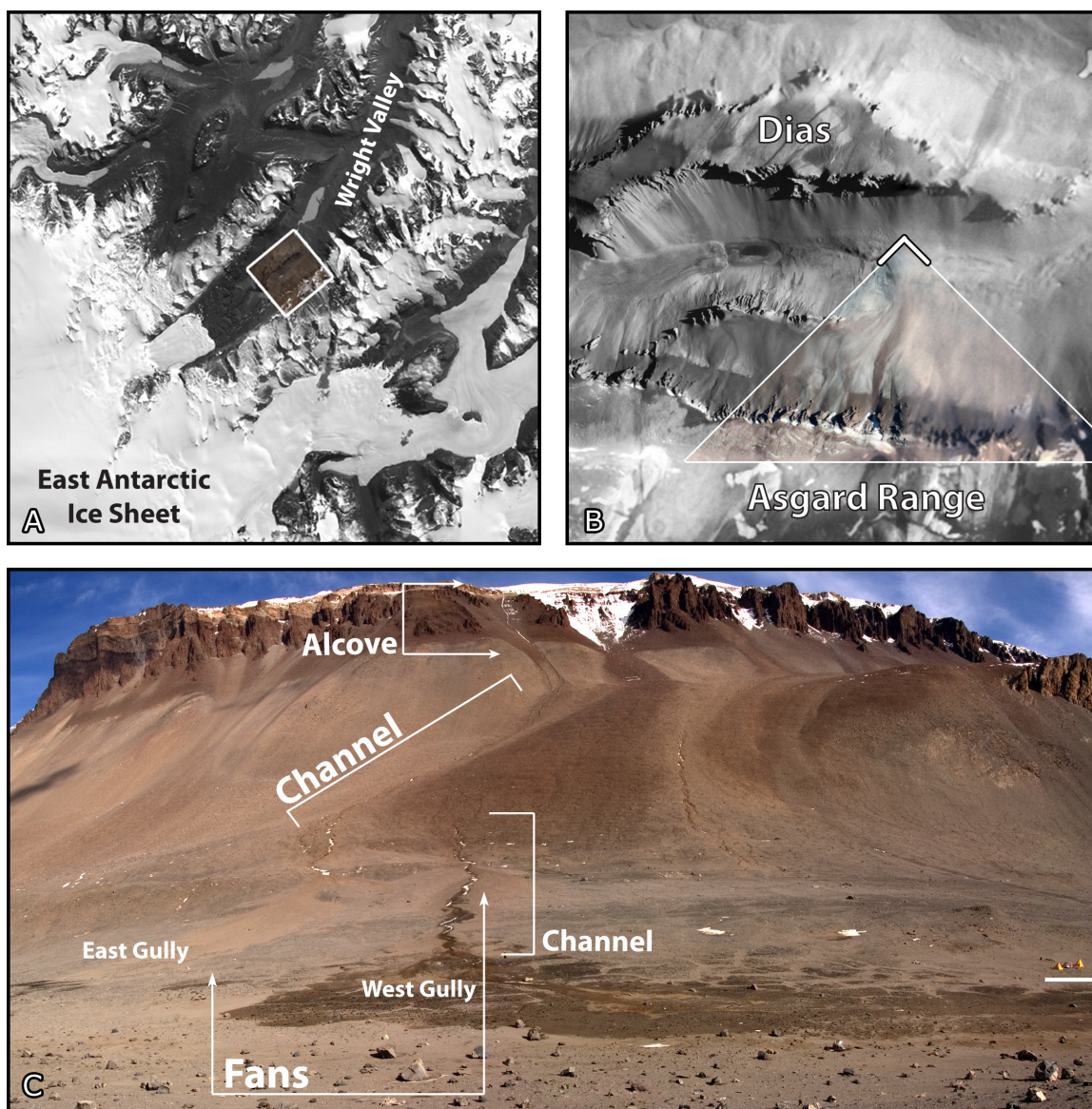
Fig. 13. Meteorological differences between the alcoves and valley floor. A. The alcoves can be seen along the edge of the Asgard range, situated at the most elevated portion of the study site. The alcoves are 1 km above the valley floor. B. During storm events low clouds hugged the alcoves. This is due to the adiabatic expansion of moist air associated with the increase elevation. C. The clouds preferentially precipitated snow within the alcoves, which contributed to the formation of the perennial snow deposits.

Fig. 14. Alcove of the east gully. A. Satellite image of the upper portion of the east gully channel and Alcove. There is a ~500 m-wide perennial snowbank in the alcove, that was the source of meltwater. Red box shows the position from which the image in B was taken. B. Small channel connecting the alcove snowbank with the main section of the gully channel. Flow was observed in the channel during periods of peak insolation. C. Sketch map of B. highlighting the main elements. Image in A, Ikonos image: po_105577

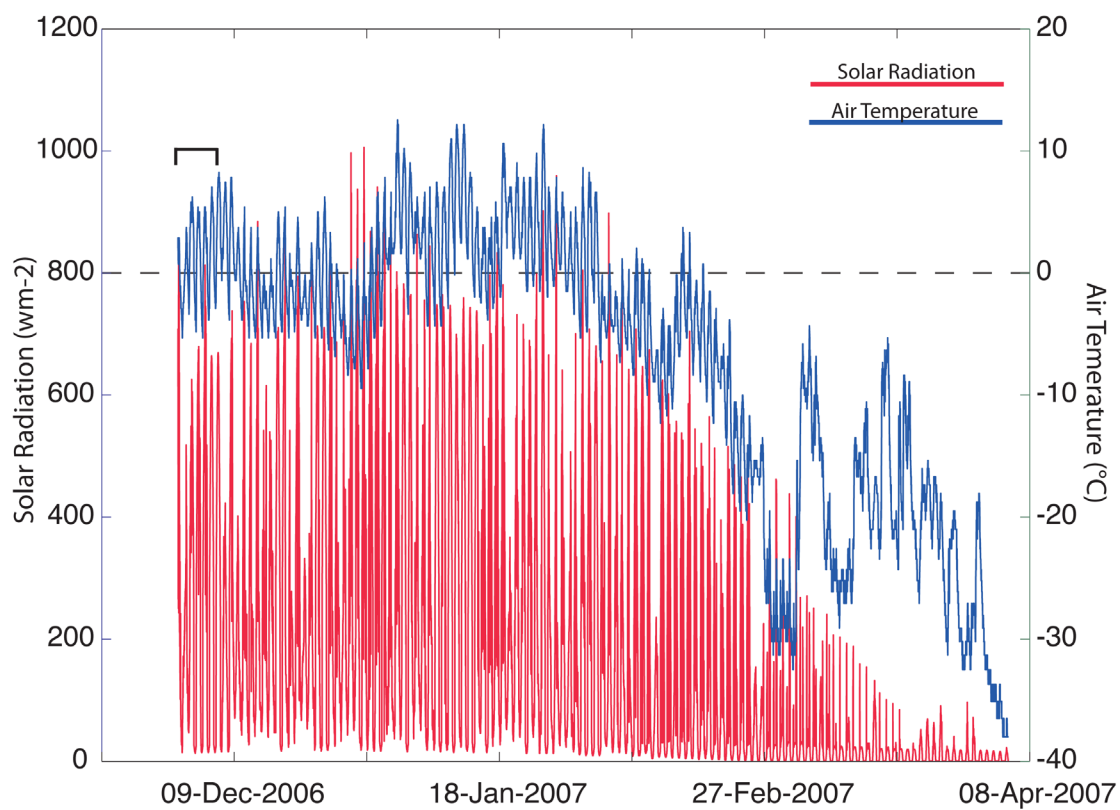
Fig. 15. Comparison between a martian gully (A,B) and the eastern gully system studied in the Dry Valleys (C,D). Both gully systems conform to the Malin and Edgett (2000) three morphological component definition of a gully (i.e. consist of a alcove, channel and depositional fan). The gullies also share morphological similarities at HiRISE full resolution scale (~0.3 m/pixel) (B,D), including the occurrence of small fluvial-like internal channels (IC) within the broader main channels (MC). This demonstrates that the Mars gullies were likely to have been eroded by fluvial erosional processes similar to the snowmelt fluvial activity that formed the South Fork gullies. Image in A and B, HiRISE: PSP_004091_1325.



Chapter 3, Figure. 1



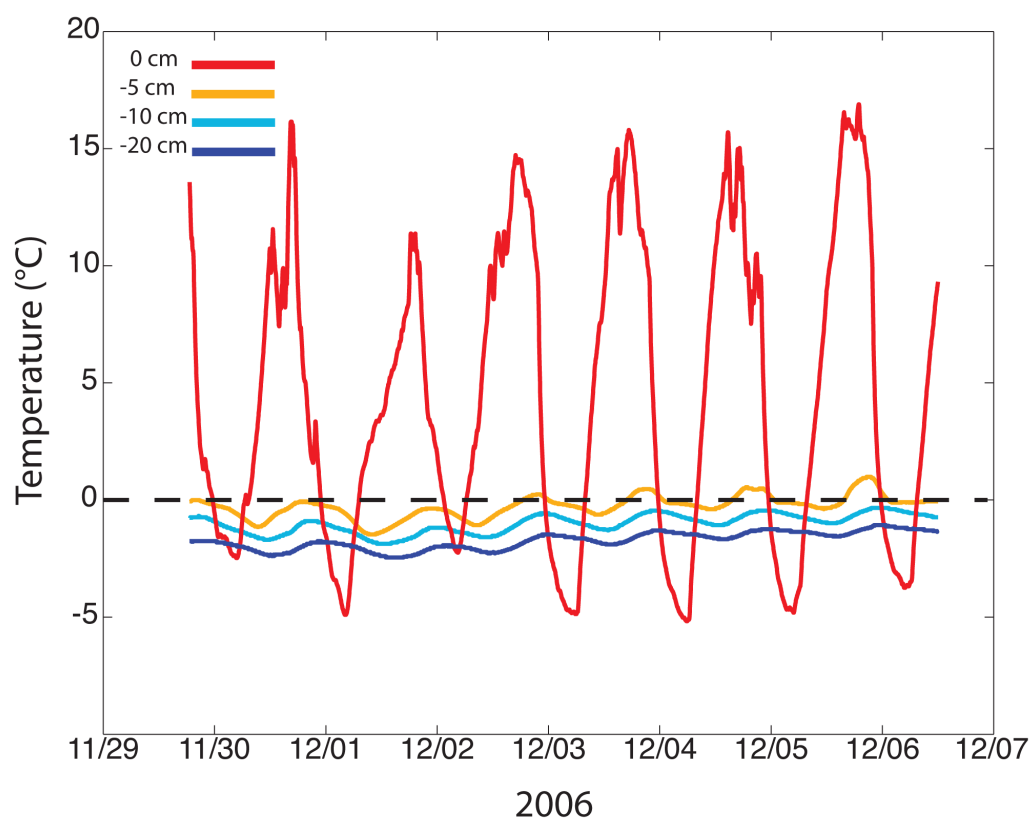
Chapter 3, Figure. 2



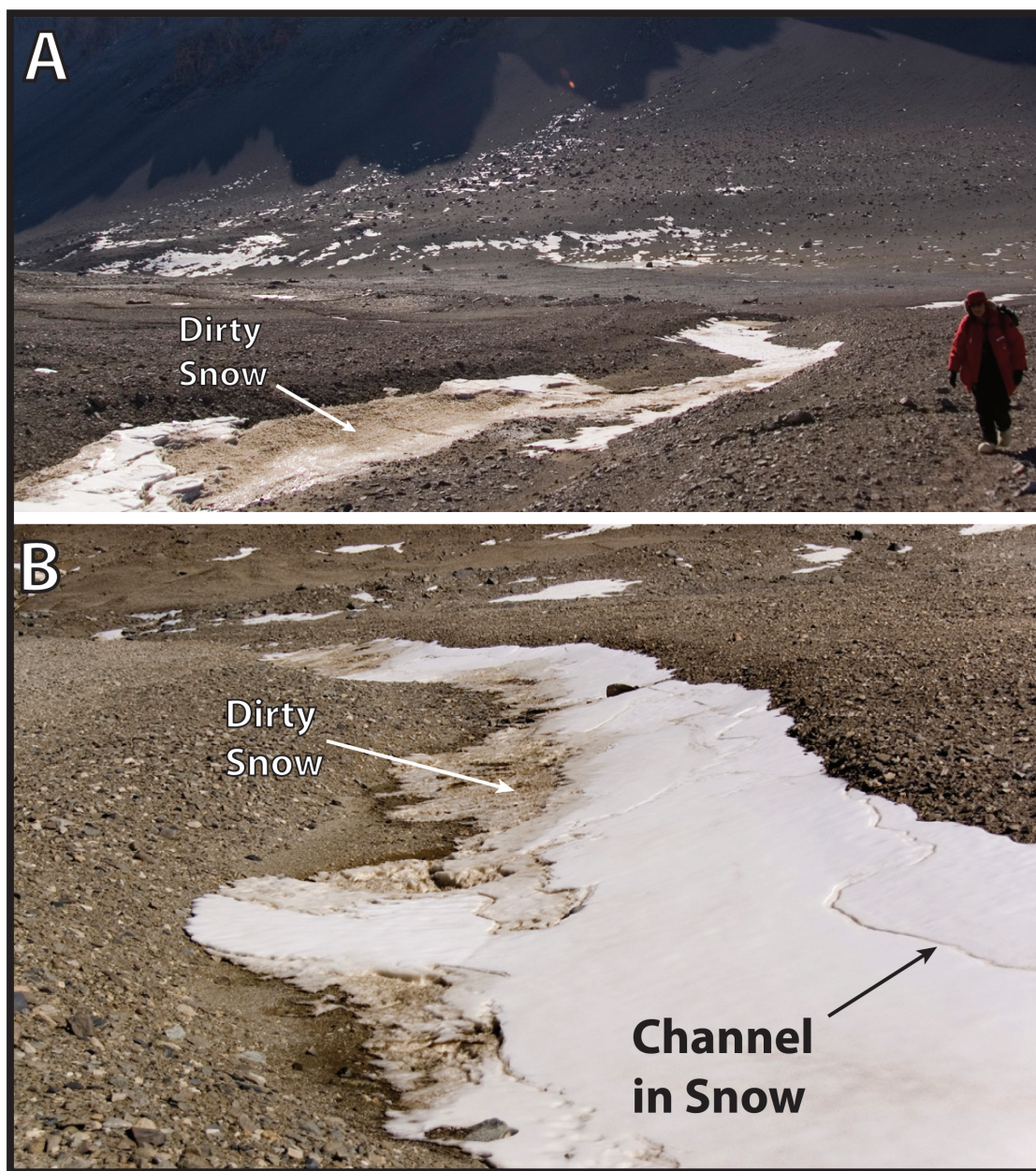
Chapter 3, Figure. 3



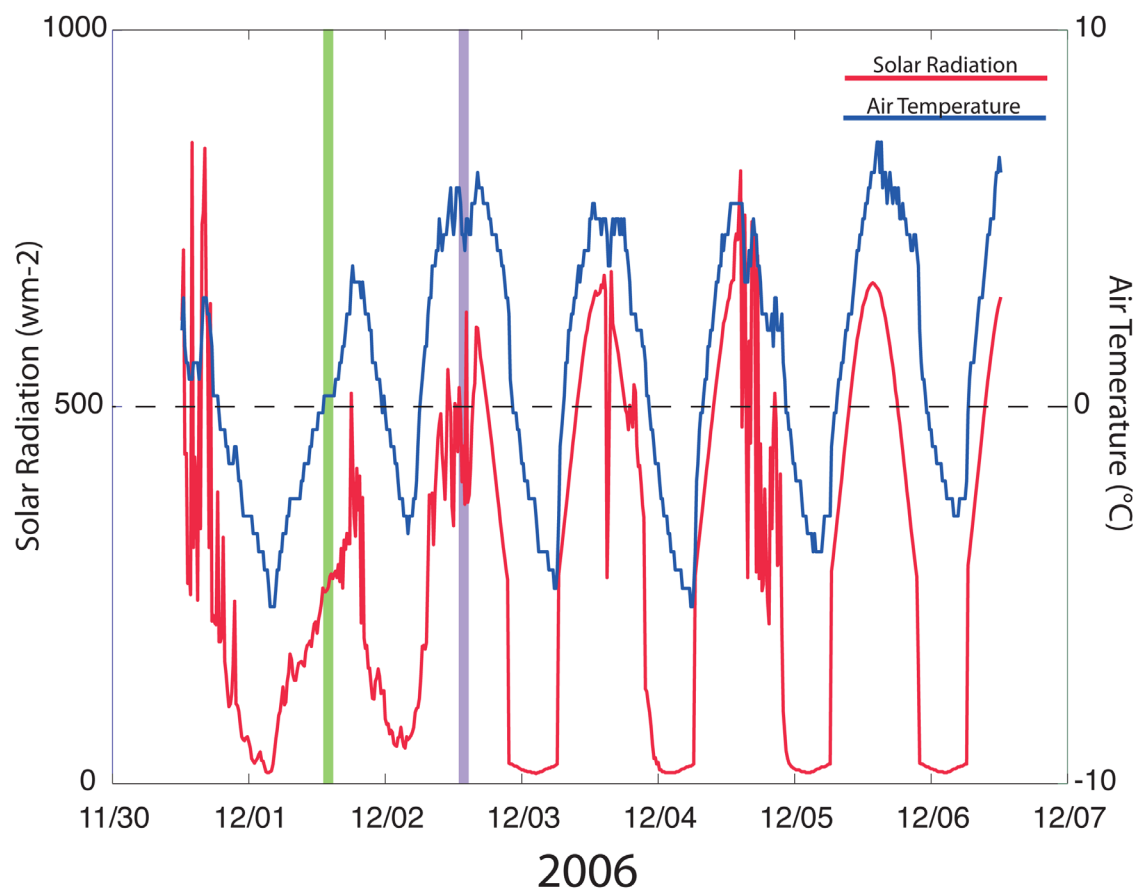
Chapter 3, Figure. 4



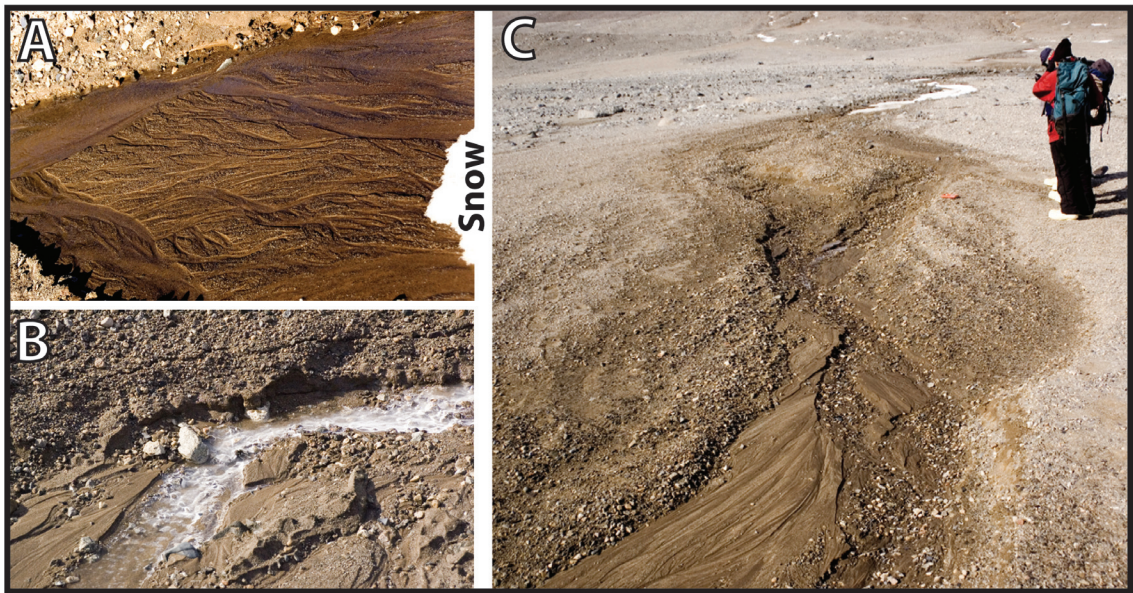
Chapter 3, Figure. 5



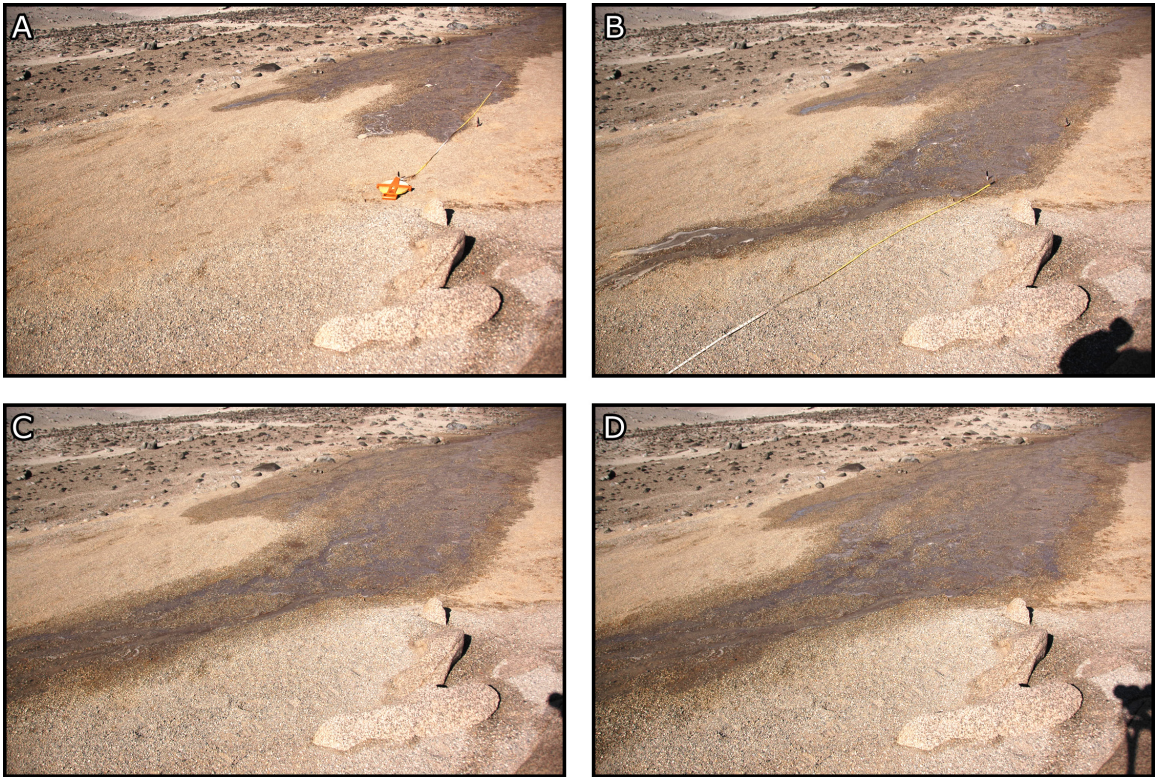
Chapter 3, Figure. 6



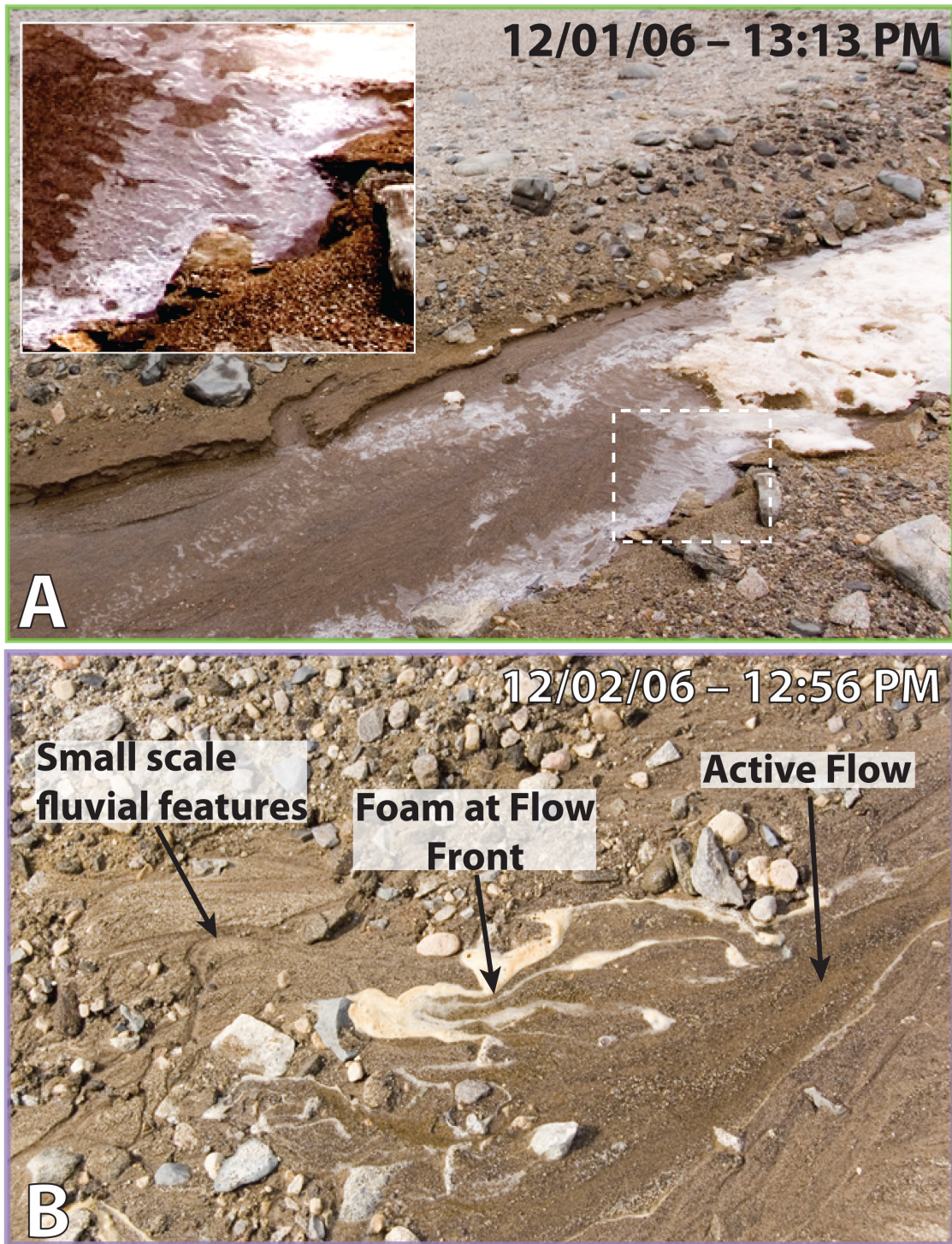
Chapter 3, Figure. 7



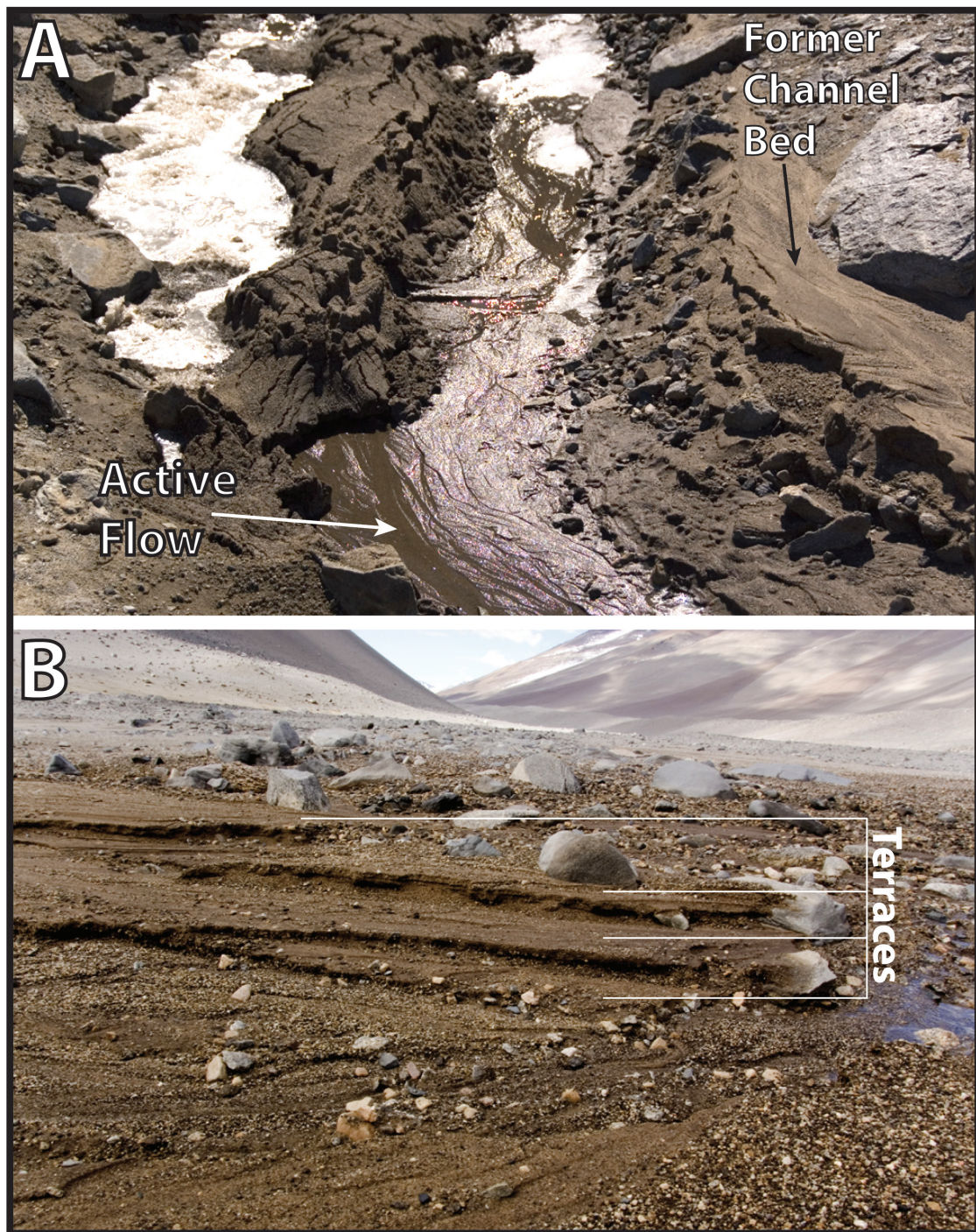
Chapter 3, Figure. 8



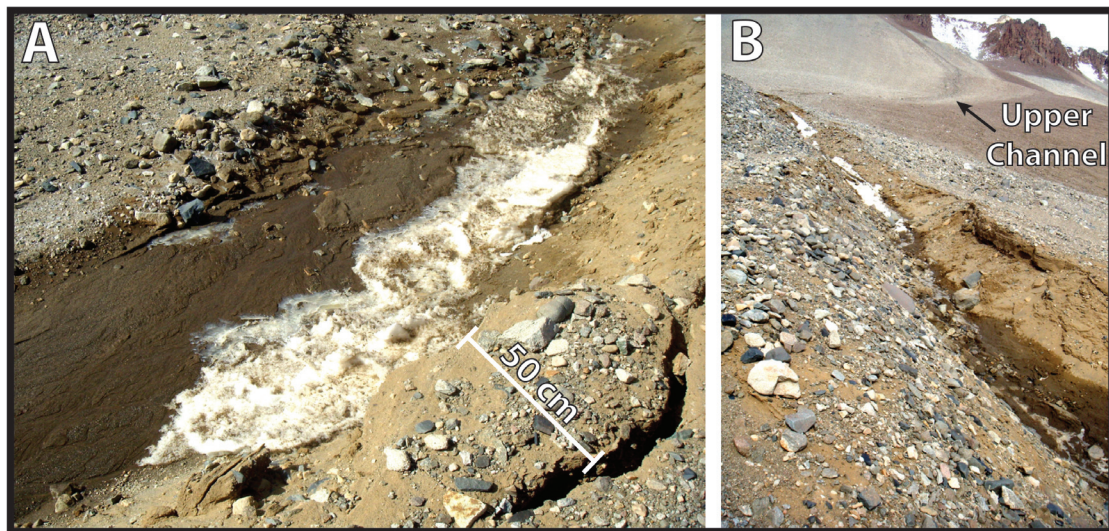
Chapter 3, Figure. 9



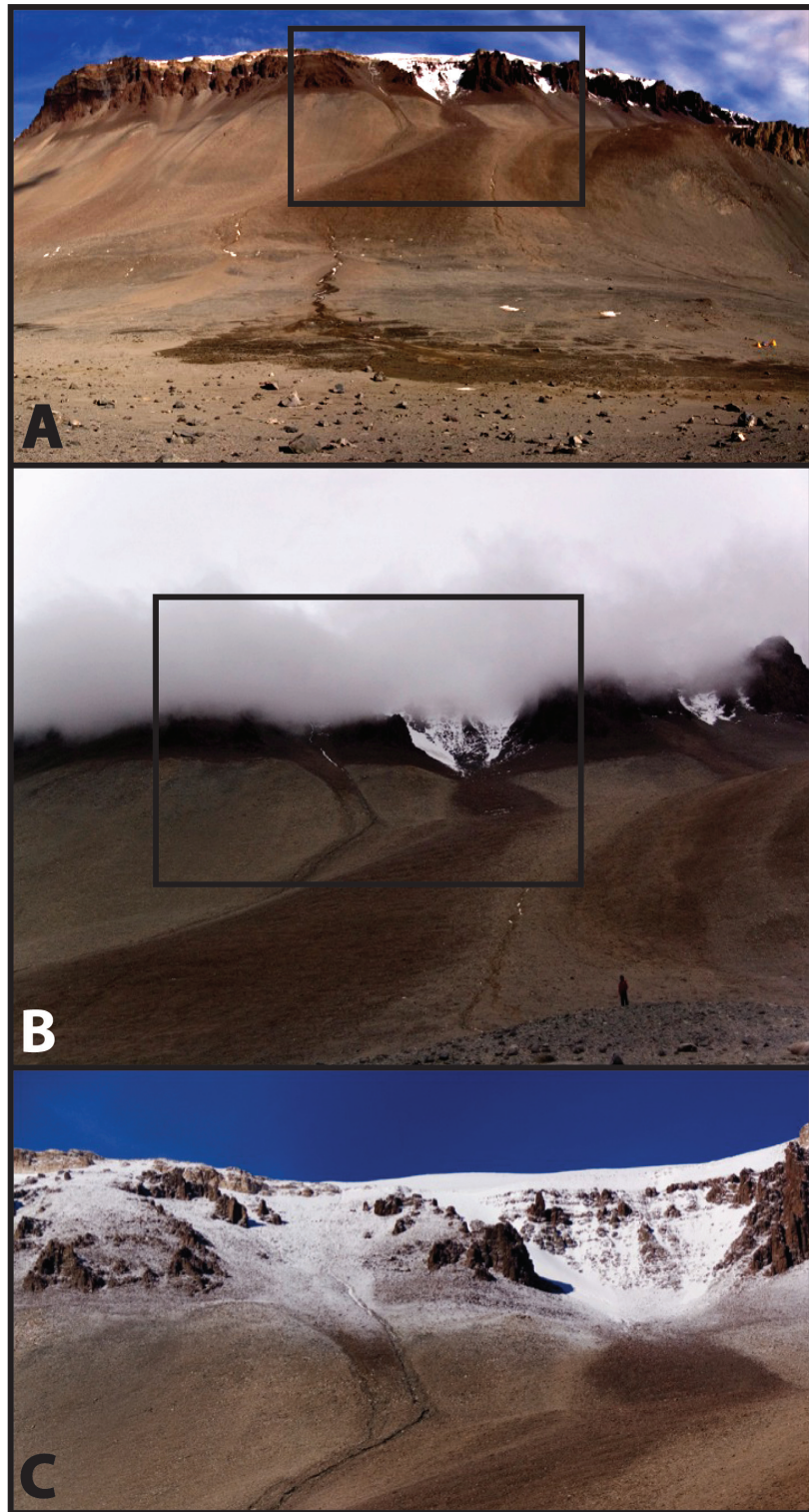
Chapter 3, Figure. 10



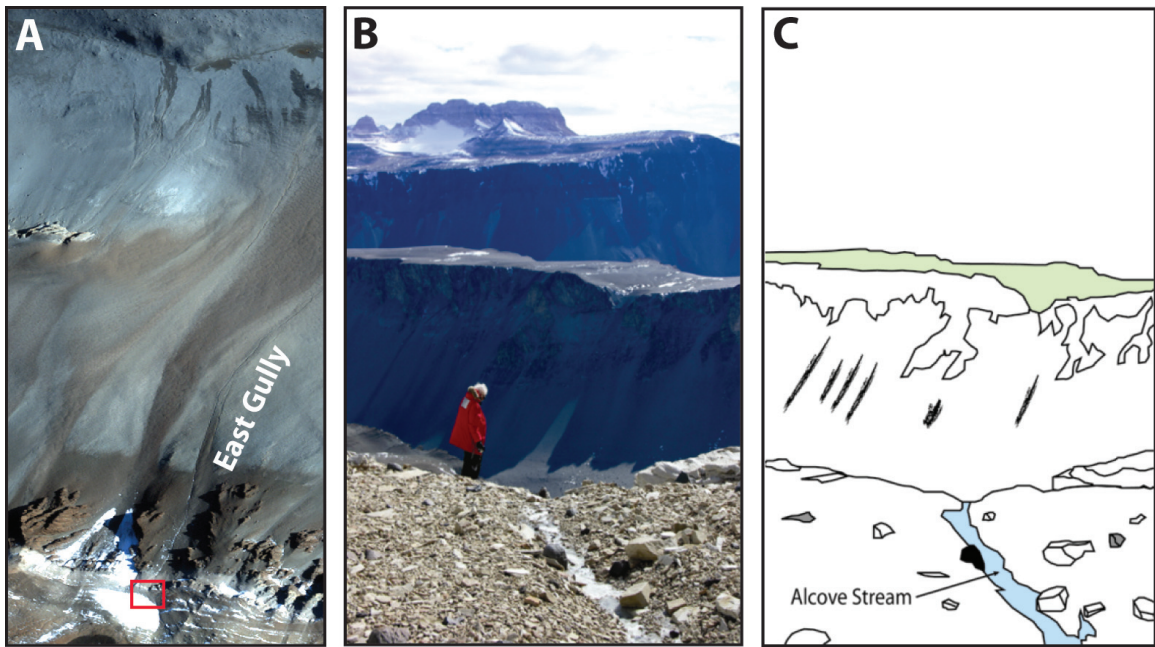
Chapter 3, Figure. 11



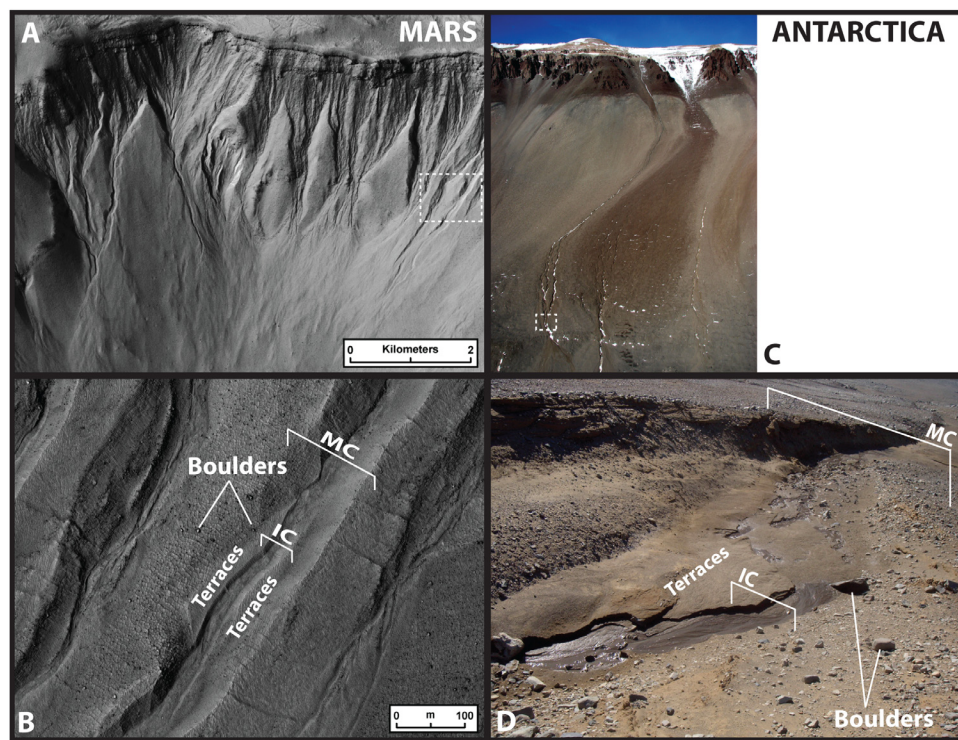
Chapter 3, Figure. 12



Chapter 3, Figure. 13



Chapter 3, Figure. 14



Chapter 3, Figure. 15

Chapter Four

The Role of Variable Solar Insolation and Topography in Recent Gully Activity in Upper Wright Valley, Antarctic Dry Valleys: Implications for Mars.

Gareth A. Morgan

Abstract

The McMurdo Dry Valleys are a hyperarid cold polar desert in which sublimation is generally higher than the small amount of snowfall that is precipitated each year. Nevertheless, annually accumulated wind blown deposits of snow can collect in topographic hollows and melt under peak daytime insolation conditions during the austral summer, providing localized surface runoff. The resulting geomorphic work of the meltwater generates small-scale gullies that are similar to young landforms identified on Mars. The combination of high topographic relief and 24 hour sunlight provides a unique natural laboratory in which to study the relationship between insolation geometry and snowmelt in a cold desert environment. We coupled a high resolution DTM and an area solar radiation model to support field observations conducted of Antarctic gullies. Our results show that the spatial distribution of gullies and the level and timing of activity observed within them is highly correlated with spatial and temporal patterns of solar radiation. This demonstrates that their formation and continued activity results from the top down insolation-controlled melting of surficial deposits of snow and ice. The sensitivity of the Antarctic gully formation to specific insolation conditions suggests that gully formation on Mars may also be highly dependent on specific solar radiation conditions and is supported by global surveys that have shown a strong aspect dependence of martian gully distribution.

1. Introduction

The McMurdo Dry Valleys are situated within the central Transantarctic Mountains, between the East Antarctic Ice Sheet and the Ross Sea and represent the largest ice-free region of land on the Antarctic continent ($\sim 4000 \text{ km}^2$, Priscu, 1998). The Dry Valleys are a hyperarid cold polar desert (Marchant and Head, 2007), characterized by minimal precipitation ($< 10 \text{ cm a}^{-1}$, Bromley, 1985), and average summer temperatures below the freezing point of water (Fig. 1c).

These conditions place significant restrictions on the hydrological systems of the Dry Valleys, and limit surface runoff to localized environments, where it occurs for short periods of the austral summer. During this time sunlight is continuous, but the high topographic relief creates a highly variable insolation environment in both spatial and temporal dimensions. The majority of overland flow and associated fluvial activity takes place within stream systems that drain into lakes that occupy portions of the valley floors (Chinn, 1993). The streams are sourced by meltwater from cold-based glaciers, which melt during periods of direct solar radiation incident on the glacier faces (Fountain et al., 1998). In areas where alpine glaciers are absent, melt from annual and perennial snow accumulation sources has been observed and documented as a source of fluvial activity within smaller scale, gully systems (Morgan et al., 2009a). Snowpack melting producing channelized flow within the gullies is correlated with periods of peak insolation during cloud-free conditions, further demonstrating the dependence of Dry Valley hydrological systems on the availability of solar radiation for the production of melt of surficial snow/ice deposits.

The gully systems present in the Dry Valleys are morphologically very similar to youthful gully features observed on Mars in high-resolution images (Malin and Edgett, 2000) (Fig 1). Gullies on Mars are defined by a tripartite morphological classification system consisting of an alcove, channel(s) and depositional fan (Malin and Edgett, 2000). The Dry Valleys gullies also exhibit these units (Fig.1), and thus provide intriguing analog landforms. The discovery of gullies on Mars has been a source of immense scientific interest because their presence suggests liquid water has flowed on the surface in the very recent geologic past, under conditions previously considered incompatible with melting and surface runoff due to pressure-temperature conditions below the triple point of water (e.g. Carr, 1996). Global surveys of the spatial distribution of gullies on Mars have demonstrated a strong correlation with latitude (restricted to latitudes above 30° in both hemispheres) and a secondary correlation with aspect (Malin and Edgett, 2000; Costard et al., 2002; Heldmann and Mellon, 2004; Dickson et al., 2007a). This demonstrates the potential for the strong influence of solar geometry on gully location and thus suggests that

martian gullies result from the melting of surface/near-surface snow/ice (Costard et al., 2002; Dickson et al., 2007a, Head et al., 2008).

In terms of precipitation, mean annual temperature and geological setting, the Dry Valleys have long been considered the most Mars-like of terrestrial environments (e.g. Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005; Baker, 2001). Therefore, the Dry Valleys provide a unique natural laboratory in which to investigate the mechanics of insolation controlled snowmelt production and gully activity and formation.

The melting of snow (S_m) is governed by the energy exchanges between the snowbank and its surrounding environment and can be expressed as (from Dingman, 2002):

$$S_m = K + L + H + LE + G + R \quad (1)$$

Where K is the shortwave (solar radiation) input, L is the longwave radiation exchange; H and LE are the turbulent exchange of sensible and latent heat (respectively) with the atmosphere; G is the conductive exchange of sensible heat with the ground; and R is the heat input by rain, which can be ignored for the Dry Valleys. The symbols used above represent the net rates of energy fluxes (energy per unit length per time) of each process (Dingman, 2002). With regards to the polar desert environment of the Dry Valleys, the most important element controlling snowmelt is solar radiation as it supplies shortwave energy directly to the snow bank (K) in addition to heating up the air and ground surface that surround the snow bank and so also influence L , H , LE and G (See Dingman, 2002, for a detailed account of the physics of snow melt). Hence characterizing the insolation environment of South Fork is essential in order to gain an understanding of gully activity and formation through the generation of snowmelt.

Integrating the effects of insolation and topography over the whole of the Dry Valleys was pioneered by the work of Dana et al. (1998). Their results showed that the solar radiation environment was highly variable due to a combination of the complex topography, coastal

cloudiness and orographic effects. We follow the broad approach of Dana et al (1998) and apply it specifically to investigate snowmelt conditions in gully systems in the South Fork of Upper Wright Valley (Fig. 2). This is achieved through the use of an area solar radiation model coupled with a high resolution (1 m/pixel) LIDAR DTM of South Fork (Fig. 2c). In addition to the modeling, in situ measurements of insolation were collected and field research on gully systems was conducted within Upper Wright Valley during the austral summer of 2006-7. Here we expand upon the results of the field research previously presented in Morgan et al (2009a) in order to investigate the relationship between gully activity and insolation and to assess the relative effects that other environmental factors have had on determining gully formation in South Fork.

2. Study Region and Associated Gully Systems

The climate of the Dry Valleys is characterized by a prominent climatic gradient resulting from a decrease in moisture and temperatures as a function of distance inland from the Ross Sea and with the corresponding increase in elevation toward the East Antarctic Ice Sheet. Marchant and Head (2007) subdivided the Dry Valleys into three microenvironments, each of which has distinctive geomorphic characteristics that reflect the availability and volume of liquid water associated with this gradient (Fig. 1c). The coastal microenvironment zone is the only region of the Dry Valleys to experience wide scale ground thaw due to summer subsurface soil temperatures exceeding 0°C and reaching the top of the ice table.. This generates a wet active layer and can cause the formation of ice wedge thermal contraction crack polygons. The formation of liquid water is highly localized in the majority of the Dry Valleys and only negligible quantities of snowmelt occur within the most elevated regions that comprise the stable upland zone; the low summer temperatures resulting from the high altitude of the stable upland zone prevents the development of an active layer. A dry "active" layer occurs in the intermediate

zone between the coast and the most elevated regions, although meltwater production and runoff in the form of streams and gullies does occur within localized areas.

In order to investigate the conditions most applicable to martian conditions we concentrated our efforts within the South Fork region of upper Wright Valley (Fig. 2). This represents the most elevated (and hence driest) portion of the intermediate microclimate zone, where it borders the colder and more Mars-like stable upland zone. South Fork consists of a 2 km wide valley formed between the northern slopes of the Asgard Range and the Dias. Local relief is > 1 km. This region marks the most inland extent of active fluvial features within the Dry Valleys and is most analogous to Mars during the Amazonian in terms of precipitation and maximum surface temperature values (Fig. 1c). Unlike the Dry Valley streams, there are no significant adjacent glacial systems to supply meltwater to the South Fork gullies (Fig. 3, 4), and so the most significant ice deposits within the study region, as established by long-term photographic records, are annual and perennial snowbanks. The effect of highly variable solar altitude and azimuth angles due to the high latitude and the extreme topographic relief of South Fork results in complex spatial and temporal patterns of insolation (Dana et al., 1998). Thus, the combination of the terrain and the extreme aridity presented optimum conditions to study and explore the relationship between insolation geometry and snowmelt.

In addition to gully systems, South Fork is also host to several ponds and a 2 km long debris tongue that occupies the western half of the valley floor (Fig 2, 4a). The most significant and only true perennial body of surface water is Don Juan Pond, which is situated in the center of South Fork (Fig. 2). Don Juan Pond is filled with a hypersaline calcium chloride solution, and remains unfrozen throughout the year, even down to winter temperatures of -50°C (Chinn, 1993). This site is thus of special scientific interest, and so understanding the potential contribution of the gullies to the hydrological system in which Don Juan Pond is situated is of value. The debris tongue terminates along the western shore of Don Juan Pond and has been broadly described in the literature as a ‘rock glacier’ (Hassinger and Mayewski, 1983; Bockheim and McLeod, 2008).

Though little specific research has been devoted to the composition and formation of the feature. Morgan et al (2009b), supported earlier work that concluded that the tongue is a relict feature (Hassinger and Mayewski, 1983) and went on to argue that the gullies are actively incising and degrading the feature (Fig. 4a).

Morgan et al (2009a) reported on the detailed investigation of two gully systems situated along the southern valley wall of South Fork to the east of Don Juan pond (Fig. 2, 3). These gullies were observed to be sourced by the melting of snow from two main deposits: 1) Annual deposits within the gully channels. Despite the low precipitation rates, strong winds redistribute snow and deposit it within topographic traps such as the gully channels themselves, permitting the accumulation of snow banks > 40 cm thick. 2) Perennial snowbanks within gully alcoves. Multiyear aerial photographs revealed the long-term stability of large (>100m) snowbanks within the alcoves situated along the edge of the Asgard Range, 1000 m above the floor of South Fork. Meltwater generation and associated fluvial activity occurred as the result of the melt of both types of snow deposits due to surface heating during periods of direct solar radiation. Below we extend the field documentation of gully systems initially reported in Morgan et al (2009a) to include the whole of South Fork.

Fifteen gully systems were identified in South Fork and they were found to display a range of channel morphology and scale. During the 2006–2007 field season, approximately one-half of these gullies were observed to display some form of surface runoff within their channels. However, interannual activity within Dry Valley streams and gullies has been noted (Conovitz et al., 1998; McKnight et al., 1999; Morgan et al., 2009c) and thus, we are cautious not to conclude that non-active gullies are much older relict features. The majority of gullies are located along the southern valley wall of South Fork (Fig. 2, 3, 4b) and consist of channels 1–2 m wide. Some display prominent levees along the channel flanks that are 0.5–1 m wide (Fig. 4b). The largest gully system in terms of channel width (> 4 m) is present along the surface of the debris tongue and consists of a multiple tributary system that feeds a central channel that is incised into the

snout of the tongue (Fig. 2, 4a). This channel opens up into a small delta close to the shore of Don Juan Pond. A solitary gully system is also present along the northern wall of South Fork on the flanks of the Dias (Fig. 4c). The gully channel is narrower (~ 1 m) than most of the other gullies, although annual deposits of snow were trapped within it (Fig. 4 c), in a manner similar to what was observed elsewhere in South Fork.

3. Methodology

3.1 Solar Radiation Model

The solar radiation analysis tools package of ESRI ArcMap GIS 9.2 software suite was used to calculate the variation in insolation across the surface of South Fork for the solar azimuth and altitude angles corresponding to early December, the period of observed gully activity). This model is based on methods from the hemispherical viewshed algorithm developed by Rich et al. (1994), and as further developed by Fu and Rich (2000; 2002) and Fu (2000). The model calculates the total solar radiation as the sum of the direct and diffuse radiation received at each point on the topographic surface. Direct radiation for each pixel of the DTM is estimated through the modeling of individual sun maps, which are a hemispherical projection of the solar trajectory for the time period over which the model is run. This takes into consideration atmospheric effects on insolation resulting from the transmissivity of the atmosphere, and the relative optical path length; the latter is a function of elevation, as it decreases with atmospheric pressure, resulting in higher irradiance at greater elevations (Dana et al., 1998). Topography is accounted for through a calculation of the slope of each model pixel derived from the DTM. Diffuse radiation is calculated using the same hemispherical projection and takes into account local slope. The effect of topographic shielding is accounted for by the estimation of view shed for each pixel. A view shed is a hemispherical representation of the topographic obstruction of the sky caused by the surrounding terrain, and was interpolated from measurements in 32 directions from each DTM

cell. Due to the high relief ($> 1\text{ km}$) of South Fork, topographic shielding was predicted to have a significant effect on the insolation environment. The contribution of reflected solar radiation is not included in the calculation of total radiation by this model, though its contribution is only considered to be minimal. The effect of clouds was also emitted from our simulations, as we were interested in investigating radiation environment under peak conditions, to assess the maximum potential for melting to occur. Previous studies have highlighted the high occurrence of cloudy conditions in the Dry Valleys (e.g. Linder et al., 1994; Dana et al., 1998) and specifically in South Fork (Morgan et al., 2009a). However, clouds are caused by both regional weather systems and orographic effects and therefore, are highly variable over space and time (Dana et al., 1998; Morgan et al., 2009a). We used pyranometers (light sensor with a hemispherical view that measures total insolation: direct + diffuse solar radiation) in the field (see below) to measure the effect of clouds on solar radiation, which was then directly compared with the modeled peak insolation conditions.

The DTM used as the base for the solar radiation models was derived from the USGS airborne LIDAR measurements of Wright Valley (2 m resolution, NSF/NASA/USGS with processing by Schenck et al. [2004]), cropped to include the South Fork study area. This data set provides unprecedented detail of the topography of South Fork and was of sufficient resolution such that the channels of some of the gully systems were clearly identifiable in the data. The upper portions of the northern flanks of the Asgard Range (Southern valley wall of South Fork, Fig 2c), are not covered by the LIDAR DTM. To account for this, a 200 m Radarsat orbital radar-derived DTM (Liu et al., 2001) was up-sampled to the LIDAR resolution and integrated into the South Fork topographic data to cover the gaps in the LIDAR coverage. The 200 m DTM was too coarse to permit studies of the upper portion of the Asgard Range and the gully alcoves in particular, but its inclusion was essential to accurately model the view sheds of the DTM cells along the lower slopes and valley floor.

3.2 *In situ* Data Sets

Silicon pyranometers were installed within South Fork to compare observed insolation values directly with the modeled data. The pyranometers (S-LIB-M003) have a measurement range of 0 to 1280 Wm⁻² over a spectral range of 300 to 1100 nm. They were connected to digital data loggers as part of an instrumental array that contained air and surface temperature sensors, and thus provided measurements of the warming effect of the sun over the study period. The sensors were concentrated along the two gully systems documented in Morgan et al (2009a), as these exhibited some of the highest levels of gully activity and were studied in the most detail. One station was established along the lower portions of the west gully channel (Fig. 2b, 3), where fluvial activity occurred as the result of the melting of annual snowpacks. The second station was positioned within the gully alcove (1000 m above the lower sensor), adjacent to a 500 m wide perennial snowbank that was observed to supply the upper portions of the east gully channel with meltwater (Fig. 2b). The insolation measurements from these stations were directly compared with the modeled results.

4. Results and Discussion

The first stage of estimating solar radiation taken by the model is to determine the solar trajectory in the sky relative to a point on the ground, for the time of year and day of interest. The altitude (a ; elevation of the sun above the horizon) and azimuth (α , angle between the sun and true north) of the sun's position within a hemispherical projection of the sky is calculated through the use of spherical trigonometry and provides the following relationships between the two angles as presented by Gates (1980):

$$\sin a = \sin \phi \sin \delta + \cos \phi \cos \delta \cos h \quad (2)$$

$$\sin \alpha = -\cos \delta \sin h / \cos a \quad (3)$$

where ϕ is the latitude of the observer, δ is the declination of the sun (related to the time of year as determined by the Earth's obliquity) and h is the hour angle of the sun. Hence the altitude of the sun is a function of latitude and the time of year and day (Gates, 1980). Maximum direct insolation on a given point of the Earth's surface will occur at midday (true solar time) on the summer solstice, when the declination of the sun is positive and at its greatest value (equal to the Earth's obliquity: $\sim 23^\circ$). This occurs in the southern hemisphere about the 21st of December.

Gully activity within South Fork was observed prior to the solstice and only occurred during the first two weeks of December 2006 (with the exception of the gully system on the lobe tongue, Fig. 4a). This indicates that insolation was sufficiently high prior to the solstice to permit the melting of snow. Despite the availability of annual deposits of snow within the gully alcoves, the majority of activity resulted from the melting of annual snowbanks of wind-blown origin, within the gully channels, and thus the volume of snow available in these deposits determined the duration of gully activity. Our study concentrates on the insolation environment of South Fork during early December 2006 (Figs. 5 – 8.) in order to capture the period of maximum activity.

Figure 5 shows a hemispherical view of the path of the sun in the sky, as it would be observed from an unobstructed viewpoint within South Fork for the 5th of December. At this time of year, daylight is continuous over a 24 hour period, but the altitude of the sun varies significantly, between 9° and 35° , with the latter (higher) value corresponding to the period of highest insolation occurring over a full day. Hence the availability of solar radiation for meltwater production is a function of the time of day (azimuth). However, the intensity of insolation on the Earth's surface is also related to topography and will be determined by slope, aspect, altitude, topographic shielding and meteorological effects associated with elevation (orographic effects). Below we discuss how the extreme topography of the South Fork of Upper Wright Valley (Fig. 2) influences this diurnal variation of solar radiation.

4.1 Terrain Effects

The high valley walls (> 500 m) and steep ($> 25^\circ$) slopes (Fig. 2) generate topographic configurations that optimize a wide range of insolation conditions and produces non-negligible topographic shielding effects that have a significant effect on the gully systems situated within South Fork. The effect of relief on insolation is demonstrated in Figure 6; this shows the extent to which the slopes of the Asgard Range and the walls of the Dias obscure the view of the sky seen from the west gully pyranometer station (Fig. 2). In addition to restricting the influx of diffuse radiation close to the horizon, the Asgards also block out direct sunlight for ~ 8 hours of the morning and late evening. This pattern was confirmed by the pyranometer at the west gully as demonstrated by the data from the 5th of December (Fig. 7). This day was chosen because it represented a cloud-free day during the period of gully fluvial activity. During the early morning and late evening when the sun is behind the Asgard Range relative to the pyranometer, the illumination of the gullies is limited to scattered light (Fig. 6, 7). This period corresponds to the diurnal periods of minimum solar radiation ($\sim 15 \text{ Wm}^{-2}$) in Figure 7. When the sun rises above the valley walls, the insolation rapidly increases and conforms to the sinusoidal pattern defined by the relationship between the altitude of the sun and azimuth described in equations 2 and 3. The only significant discrepancy between the modeled (Fig 6) and observed (Fig. 7) results is the time that the sun rises above the Asgard Range; it is an hour earlier in the model. This is likely to be the result of the low resolution of the 200 m DTM used to model the topography of the upper slopes of the Asgard Range, due to the absence of LIDAR data coverage for this area. Nevertheless, the strong correlation between the modeled and recorded data demonstrates the ability of the model to simulate the topographic effect on the spatial variability of insolation.

Melting and the subsequent initiation of surface runoff within the two gully systems adjacent to the west gully pyranometer station occurred after the sun had risen above the Asgard Range. Runoff lasted for ~ 7 hours, a period when solar radiation was above $\sim > 400 \text{ Wm}^{-2}$. The

occurrence of discharge was also dependent on the physical properties of the channel, including depth and moisture content of the active layer, which in turn governed the ratio of infiltration to surface runoff. Solar radiation at the west gully station was also observed as the driving force behind local air and surface temperatures, both of which show strong correlation with insolation variations (Fig 8).

Figure 9 shows the results of the solar radiation model applied to the whole of South Fork for six one-hour runs corresponding to different periods over the course of the 5th of December. The location of the west gully pyranometer station is shown on the maps and can be compared with Fig 7. The model runs demonstrate how the Asgard Range casts the whole of South Fork in shadow during the middle of the night, but that insolation values rise quickly towards midday in accord with the rising solar altitude (Fig. 9). The highest values of solar radiation simulated over the entire day occur along the northern (hence pole facing) slopes at midday. During this time these slopes ($> 30^\circ$) are near orthogonal to the sun during its highest position in the sky (Fig. 6). These slopes also contain the majority of gullies in South Fork. In contrast the southern (pole facing) slopes receive significantly less solar radiation, and are illuminated only when the sun is relatively low in the sky prior to and immediately after it dips behind the Asgard Range. Gullies are absent from the northern walls (Fig. 9) and this supports the interpretation that high insolation conditions are required for gully formation in South Fork. This in turn, further supports the interpretation that gully activity is the result of top-down melting of surficial deposits of snow/ice (Head et al., 2007; Morgan et al., 2007; Dickson et al., 2007; Levy et al., 2007; Morgan et al., 2009a). The one exception to this poleward-equatorward facing asymmetry is the presence of a gully system close to the western edge of the northern wall (Fig. 2,4c,9). Close inspection of the topographic data reveals that the portion of the pole-facing slope on which the gully occurs is orientated away from true south by $>5^\circ$. As a result, unlike the rest of the slope, the gully receives direct insolation during the late morning when the sun's altitude is $\sim 30^\circ$ (Fig. 9, 10). This value is higher than the rest of the northern slopes receive, but is lower than the insolation incident on the

southern slopes at mid-day. This may explain why this gully occurs where it does, and why the gully channel is not as wide as the those on the southern slopes. Thus the topographical positioning of this gully is likely to represent the minimum solar radiation required for gully formation to occur.

The largest gully system in terms of channel width, and the most active during 2006–7, is situated along the surface of the debris tongue (Fig. 2, 4a, 9). The low surface angle (slope) of the tongue ($\sim <10^\circ$), and its position along the western portion of the valley floor, results in it receiving lower levels of insolation than the southern slopes, but significantly more than the northern slopes (Fig. 9). An explanation for the size and high levels of activity may be that it is supplied water from the equator and northeast facing slopes that receive high levels of insolation (Fig. 9). Pits dug in the active layer upslope of the gully channel became partially filled with water upon excavation. This could represent active throughflow of water above the ice table that has originated along the adjacent gullied slopes. This is consistent with soil measurements conducted by McLeod et al (2008) that suggest that the relative shallow depth of the ice table (<70 cm) along the southern wall of South Fork is the result of meltwater sourced from higher elevations that has subsequently frozen at depth. The large gully system incises the snout of the tongue and opens up into a small delta (~ 5 m across) on the valley floor (Fig. 4a). Runoff emanating from this delta flows directly into Don Juan Pond. The mapping of the gully systems challenges previous assumptions that snowmelt does not provide a significant volume of water supplying Don Juan Pond (Chinn, 1993); these models hypothesized that the melting of ice from within the tongue itself was a more significant input of water than gully runoff.

The absolute values for insolation are underestimated in the model by $\sim 200 \text{ Wm}^{-2}$ during the peak periods of insolation (12 pm, solar time, Fig. 7, 9). This is likely to be the result of the atmospheric parameters chosen as input to the model. The combination of: low moisture content (due to the desert environment), low particulate matter and aerosols (due to Antarctica's isolation from anthropogenic emissions) and low ozone can account for higher atmospheric transmissivity

and thus higher insolation values. The unusual atmospheric properties of the Dry Valleys environment needs to be taken into account when modeling snowmelt in the Dry Valleys.

4.2 Orographic Effects

Perennial snowbanks within the alcoves along the edge of the Asgard Range have been identified as a source of meltwater for some of the South Fork gully systems (Morgan et al., 2009). These alcoves are a kilometer above the floor of South Fork and as a consequence are characterized by different meteorological conditions in addition to insolation differences associated with elevation.

Solar radiation is attenuated as it passes through the atmosphere. As a consequence, the sunlight incident on the surface of the Earth will be less than that received at the top of the atmosphere because of a combination of factors contributing to the attenuation, including water, dust and ozone content (Gates, 1980), as discussed in the previous section. Therefore, under clear sky conditions insolation on a given surface will increase with elevation. Figure 11 compares the insolation measured in the alcove with that measured at the west gully station on the lower portion of the southern slopes of South Fork. Peak insolation on cloud-free days (e.g. the 5th and 6th of December) was recorded as 150 Wm⁻¹ higher in the alcove than along the lower gully channels. This difference will not be completely attributed to the lower atmospheric pressure, as the large snowbank adjacent to alcove pyranometer would have contributed a significant amount of reflected sunlight to the sensor.

Morgan et al (2009a) discussed how the effects of the adiabatic lapse rate of the atmosphere (associated with the decrease in atmospheric pressure with altitude), significantly limited the amount of melting experienced by the perennial snowbank in the alcove. Figure 8 shows the air temperature recorded at the west gully station plotted against the air temperature data from the alcove of the eastern gully, 1 km above (note: both stations recorded air

temperature at exactly the same time and so each point on the plot represents one point in time). A linear fit to the data shows a -10°C offset between the lower elevated west gully station and the alcove station. This shows very good agreement with the expected dry adiabatic lapse rate (9.8°C per 1 km elevation), and is interpreted to be the reason that the air and surface temperatures in the alcove are consistently below 0°C , and thus significantly lower than the temperatures recorded along the lower gully channels (compare Figs. 11 with 12). The thinner atmosphere at this elevation thus results in the alcove snowbanks losing heat more effectively through advection. Hence lower amounts of melt occur within the alcove despite the enhanced insolation. Therefore, the effects of atmospheric pressure need to be included in modeling of snowmelt in the Dry Valleys, as absolute values of insolation alone can be misleading.

4.3 Effect of Clouds

The model did not consider the effect of clouds, but they are a common occurrence in the Dry Valleys and in reality do have an effect on the insolation reaching the surface (Dana et al., 1998). Scattered clouds will disrupt the pattern of diurnal direct solar radiation and full overcast conditions will block out direct radiation completely. Morgan et al (2009a) demonstrated the effect of clouds on gullies, and reported a lack of gully activity on overcast days. Half of all days during the austral summer are characterized by cloudy conditions (Linder et al., 1994), and during the period of gully activity in the first half of December 2006, only two days experienced no clouds (5th and 6th, Fig 8), and three days were completely overcast. Therefore, estimates of snowmelt based on modeled insolation values need to account for the presence of clouds in order to produce realistic results. This is particularly important when considering the effect of climatic change and attempting to reconstruct gully formation under paleoclimatic conditions. Figure 13 shows simulations of insolation for the latitude of South Fork (77°S) based on the simulation of Earth's orbital parameter history (i.e. obliquity, eccentricity and argument of perihelion) by Laskar

et al (2004b). Before one can use this plot to consider the potential for gully formation in the Dry Valleys over the last 6 million years, one must assess the associated feedbacks that affect cloud formation. For example higher insolation values than are experienced at present may cause greater snowmelt, but it may also reduce the seasonal sea ice cover in McMurdo Sound more rapidly and increase the moisture, which in turn could increase cloud cover.

4.4 Application to Mars

The strong correlation between the spatial distribution of gullies and the timing of their activity with the spatial and temporal distribution of insolation in South Fork demonstrates the importance of solar radiation in gully formation in a polar desert setting. The generation of surface runoff through the melting of surficial deposits of windblown snowbanks within gully systems in the Dry Valleys thus provides a compelling model for gully activity on Mars. Although the climate of South Fork provides a solid terrestrial analog in terms of temperature and potential precipitation rates for Mars at high obliquity (e.g. Baker, 2001; Mischna et al., 2003; Marchant and Head, 2007) (Fig. 1c), the Dry Valleys still represent an environment more conducive to surface water runoff and fluvial process due to surface pressures continuously above the triple point of H₂O and the daily duration of temperatures above 273 K. Thus the location of gullies on Mars should be more restricted in terms of their spatial extent relative to South Fork due to the greater limitations on areas that receive insolation conditions preferential for melt to occur on Mars (see Hecht, 2002, for a full account on the conditions necessary for melt to occur on Mars). This is in agreement with numerous global surveys of martian that have shown gullies to exhibit clear aspect dependence as a function of latitude (Costard et al., 2002; Heldmann and Mellon, 2004; Dickson et al., 2007). This is best accounted for by insolation conditions encountered during periods of high obliquity (Costard et al., 2002). Mars has experienced significantly greater orbital dynamic variations than the Earth (Laskar et al., 2004a,b) and

consequently will have experienced a more complex insolation history than the Dry Valleys (Fig. 13). High resolution DTMs of the surface of Mars could therefore be used in the same way as the were in this study. These could be combined with simulations of orbital dynamics to explore the solar radiation environment of martian gullied regions in order to identify periods in Mars' history in which gully activity could have occurred.

The isolation of Antarctica from anthropogenic emissions has resulted in the atmosphere above the Dry Valleys having a very low aerosol optical thickness (0.04), which permits a high direct beam irradiance from the sun (Dana et al., 1998). In comparison, the atmosphere of Mars is considerably more dusty, and the associated reduction of solar irradiance at the martian surface caused by the dust content of the atmosphere needs to be considered in martian models of gully activity (Note that this is a relative effect between Mars and the Earth, as the absolute difference in insolation is a function of the distance of a planet from the sun, see Fig. 13) Clouds, on the other hand, would be orders of magnitude lower on Mars relative to the conditions that were encountered in the Dry Valleys during the field season. This is perceived to be the case even under projected increases in atmospheric pressure and water vapor during the sublimation of the polar caps under high obliquity conditions on Mars.

Martian gullies typically extend over elevation differences greater than the 1km present between the alcove and fan of the east gully in South Fork (Fig 2c). However, unlike Earth, the atmospheric lapse rate on Mars is low because of the thin atmosphere (~ one hundredth of the Earth), and so there is no appreciable surface temperature gradient between an altitude difference of ~1 km. However, because the mean surface pressure on Mars is close to the triple point of water, the associated atmospheric pressure decrease with elevation is significant with regard to water stability. Thus elevation effects will also be important on Mars, but for different reasons than on Earth (e.g. Lobitz et al., 2001).

5. Conclusions

Our investigation of South Fork agrees with the results of the Dana et al. (2003) study of the whole of the Dry Valleys; they concluded that insolation is strongly affected by topography but is also influenced by orographic effects and clouds. The spatial distribution of gullies in South Fork and the level and timing of activity observed within them is highly correlated with spatial and temporal patterns of incident solar radiation. This favors the interpretation that the formation and continued activity in gullies results from the top-down, insolation-controlled melting of surficial deposits of snow and ice. The sensitivity of the South Fork gullies to insolation suggests that gully formation on Mars will also be highly dependent on solar radiation and is supported by global surveys that have shown an aspect dependence of gully distribution.

Acknowledgements

Special thanks are extended to my advisor **James W. Head** and to **David R. Marchant**, **James L. Dickson** and **Joseph S. Levy** for extensive discussions on this work during the formulation of the ideas and content of this chapter, and for their support with the fieldwork conducted in Antarctica. I look forward to additional assistance and collaboration with all of them as this contribution is prepared for publication.

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Figure Captions

Fig. 1. Comparison between a gully within Wright Valley in the Antarctic Dry Valleys (a) and gully systems on Mars (b). On both planets the gullies consist of alcoves, sinuous channels and depositional fans, which form the basis of a gully classification system for Mars (Malin and Edgett, 2000). (c) Morphogenetic regions of Earth expressed over temperature and precipitation space (from Marchant and Head, 2007: adapted from Baker [2001]). Dashed oval shows region comprising the Antarctic Dry Valleys (ADV), including the stable upland zone SUZ (shown as a black dot), the inland mixed zone IMZ, and the coastal thaw zone CTZ. South Fork is positioned within the driest portion of the IMZ (shown as red dot). Also plotted are modern Mars conditions at 30°, 50°, and 60° latitude, as well as an ancient Mars at 300 and 1000 mbar. Martian gullies are located poleward of 30°. TD: modern conditions at Taylor Dome, 35 km south-west of the ADV; EAIS (East Antarctic Ice Sheet): modern conditions at Vostok, interior

East Antarctica (78° S); LGM: conditions during the last glacial maximum (~18 ka) in interior East Antarctica. Antarctic image (a) was taken from a helicopter during the 2006 field season. Mars image (b), subset of HiRISE image: PSP_001842_1395.

Fig. 2. Context images of the study site. A. South Fork (highlighted by white box) in the context of Wright valley. B. South Fork of upper Wright Valley. Boxes highlight location of photographs in Fig. 3 and 4. Gullies are highlighted in red. The pink circle (within box 3B) corresponds to the pyranometer station adjacent to the lower channel of the west gully. The blue circle in the Asgard Range is the pyranometer station in the east gully alcove. Note it is located right next to a large-scale (~ 500 m) snowbank. Green circle, location of sun map in Fig. 6. C. Composed DTM of South Fork, comprised of LIDAR 2 meter resolution data (Schenk et al., 2004) and radar derived 200 m data (Liu et al., 2001).

Fig. 3. The two gullies investigated in depth in Morgan et al (2009a). A. Field photograph of the two gully systems (labeled East and West) taken from the floor of South Fork. Note the occurrence of snow in the alcoves and channels. The horizontal white line in the lower right corner highlights our field camp for scale. The elevation of the summit of the Asgard Range is 1000 m above the floor of South Fork. B. View of the lower channels and fans of the two gullies. The image was taken on the 29th of December, 2006, after much of the snow had melted within the channels. Progradation of the fan of the western gully by the main gully channel is evident. This is interpreted to be the result of the positive erosional feedback effect of the gully channel trapping snow, which aids the further erosion of the channel and fan deposit. The sand wedge thermal contraction crack polygons that cover the slopes adjacent to the gullies can be clearly seen, and appear to have influenced gully channel development through the creation of troughs that can concentrate drainage and flow (Levy et al., 2008; 2009). Arrow points to North.

Fig. 4. Other gully systems in South Fork. A. Large gully channel that is incised into the snout of the debris tongue (see Fig. 2 for location). B. Two gully channels along the southern wall of South Fork. The channels display levees. C. Single gully system along the north wall of South Fork. Note the gully channel contains windblown snow,

Fig. 5. Hypothetical sun map of the solar trajectory across the sky above South Fork for the 5th of December. This would be the view seen by an observer on the ground in the absence of any topography to obstruct the observer's view of the sky. Hemispherical projection; the closer that the sun's path (orange line) is to the center of the plot, the higher the sun is in the sky. The times on the sun's path refer to true solar time. Note the 5th of December was chosen because this was one of the cloud free days of the gully activity period (see Fig. 7).

Fig. 6. View shed (a view shed is a hemispherical representation of the topographic obstruction of the sky caused by the surrounding terrain) from the perspective of the pyranometer at the west gully (Fig. 2b), overlain on the sun map for the 5th of December (Fig. 5). This demonstrates that self-shadowing by the slopes of the Asgard Mountain range prevents a direct view of the sun during the morning and late evening.

Fig. 7. Insolation measured by the west gully pyranometer (Fig. 2b) for the 5th of December. Compare with Fig 6 above. The effect of the self-shielding of the slopes of the Asgard Mountain range is apparent in the corresponding low values of insolation during the morning and late evening. The blue boxes correspond to the times of the area solar radiation model run in Fig. 9. Note true solar time is off set from the local time (New Zealand time is used in the McMurdo Area), due to the 19° difference in longitude of the actual position of South Fork from that of the central meridian of UTC+12 (180°).

Fig. 8. Insolation and its corresponding effect on surface and air temperatures as observed at the west gully station (Fig. 2b) for the 5th - 13th of December. Note that only the 5th and 6th experienced cloud free conditions. The 7th and 9th were overcast days.

Fig. 9. Area solar radiation for six one-hour periods during the 5th of December (see Fig. 7, to compare with in situ measurements of insolation). During the middle of the night the entire South Fork area is cast in shadow from the Asgard Range and thus insolation levels are minimal (A, compare with Fig. 6). During the late morning (B and C) the one gully on the southern slopes (pole facing) is preferentially oriented to receive direct insolation. This orientation difference with the rest of the northern slopes appears to have been sufficient to permit gully formation there and suggests that this is the minimum amount of insolation required for gully formation. Peak insolation occurs at midday and is highest on the southern (equator facing) slopes (D). This is also where the majority of the gullies are located. During the evening (E and F) insolation begins to decrease, except for the west-facing slope of the Dias. Between 16:00 and 18:00 pm melt in the gullies was observed to cease, and in some instances runoff froze in situ.

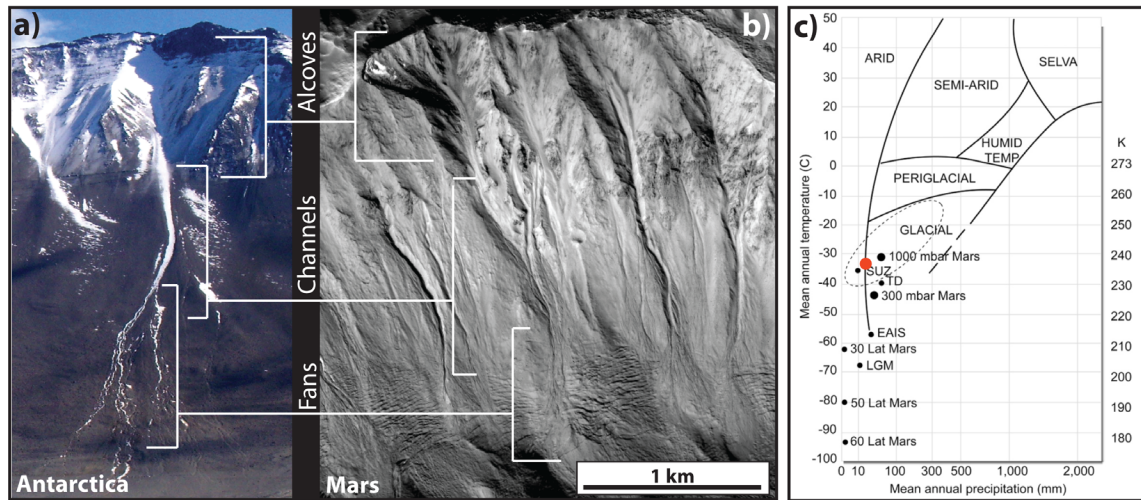
Fig. 10. View shed and corresponding sun map for the 5th of December at the position on the Dias gully highlighted with a green circle in Fig. 9. For the majority of the day the sun is blocked by the walls of the Dias, except between the hours of 3:00 – 10:00 am, when the sun reaches a maximum altitude of 30°.

Fig. 11. Insolation and its corresponding effect on surface and air temperatures as observed at the alcove of the east gully station (Fig. 2b) for the 5th - 13th of December. Insolation recorded over the same period at the west gully station has been replotted for comparison. The solar radiation values are higher in the alcove than the lower channel values; this could be 1) the

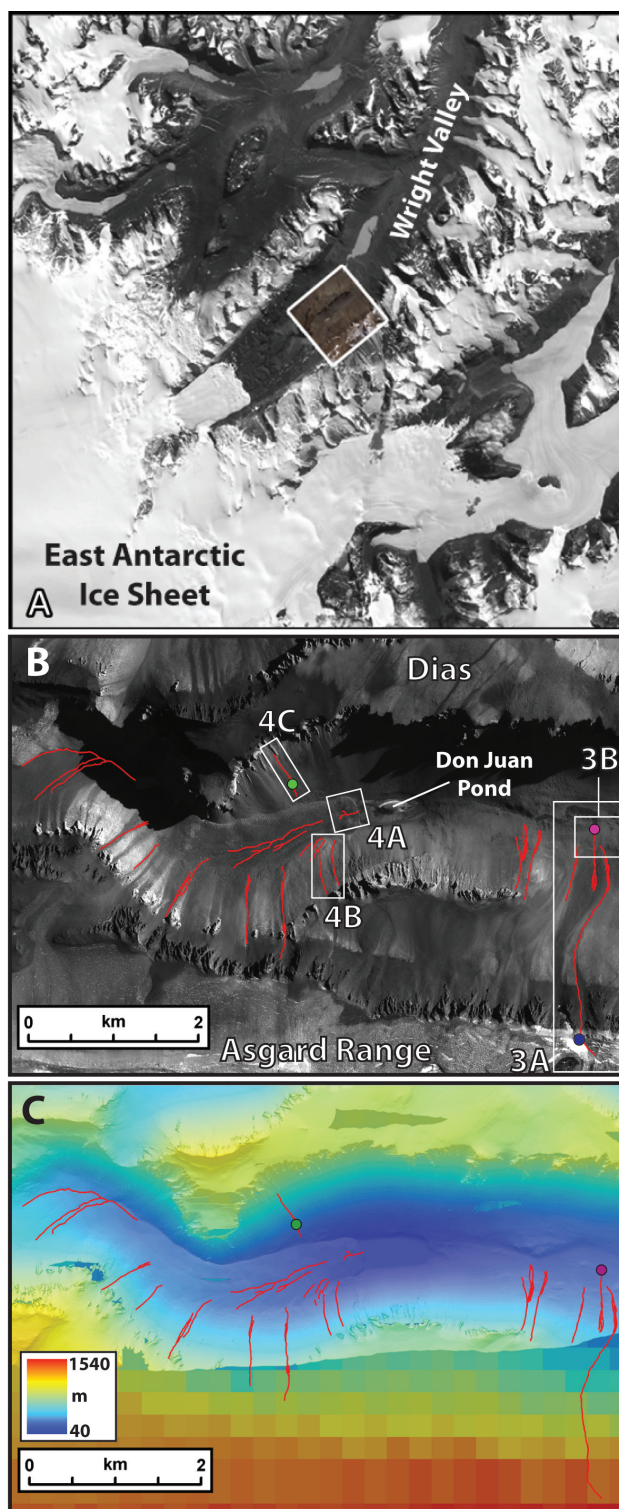
result of the lower atmospheric attenuation associated with the higher elevation of the alcove pyranometer; 2) the result of shortwave radiation reflected off the alcove snowbank; or alternatively 3) a combination of the two effects. The air and surface temperatures in the alcove are significantly colder than the west gully station values because of the atmospheric lapse rate.

Fig. 12. West gully air temperature plotted against alcove air temperature (measurements made at exactly the same time) to demonstrate the effect of the atmospheric lapse rate which has caused the alcove temperatures to be offset by -10°C relative to the west gully station, due to the 1 km difference in elevation.

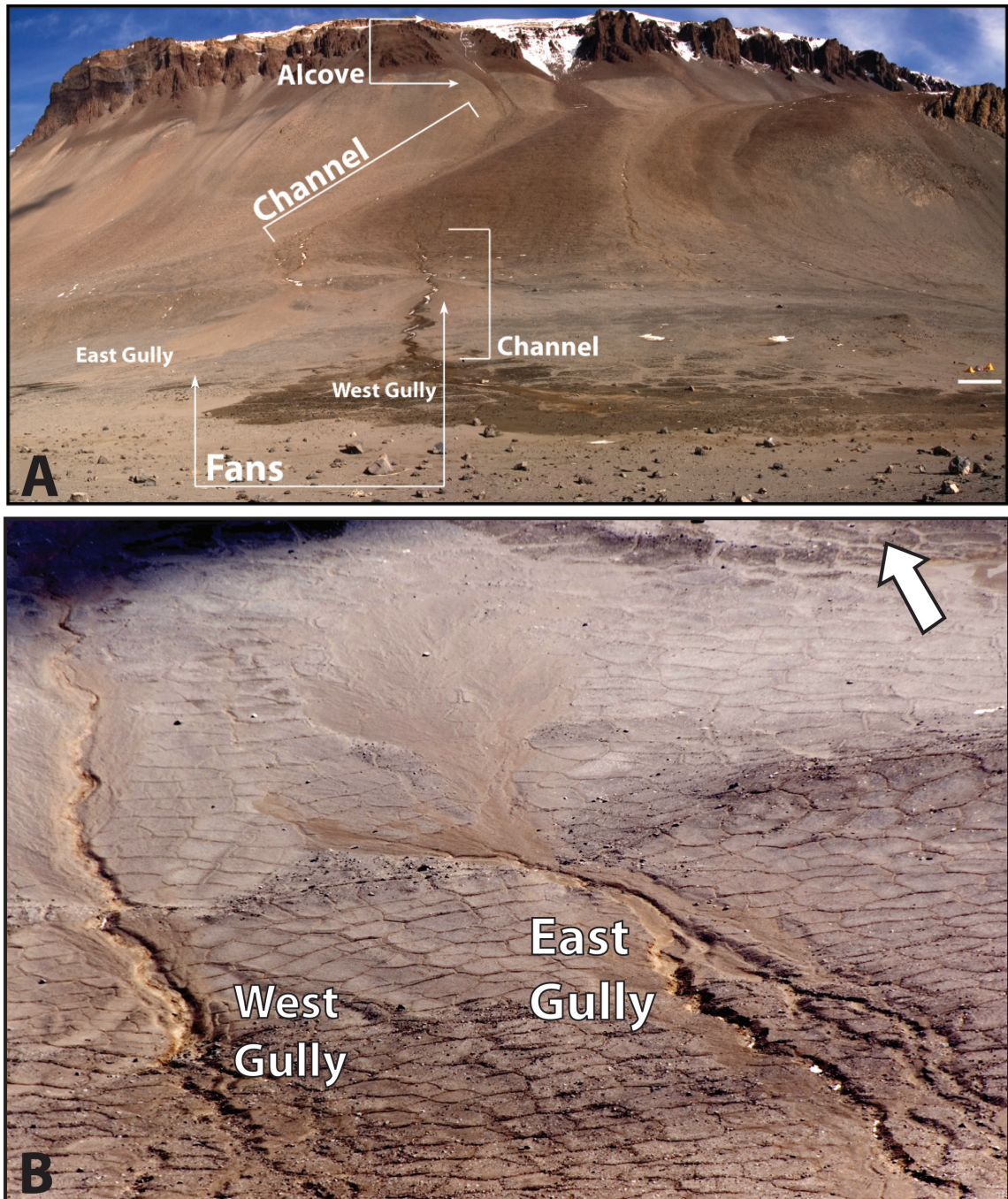
Fig. 13. Comparisons between the simulated insolation history of the Dry Valleys with that of a gullied location on Mars at 47°S over the last 6 million years. Both plots show significant variation, but the insolation of the Earth has fluctuated close to a constant value, while the Mars location shows considerable deviations over the same time period. This is due to greater range in obliquity and eccentricity values experienced by Mars relative to the Earth (Laskar et al., 2004a). Note that the Earth receives significantly more insolation than Mars due to it being closer to the Sun. Earth insolation calculated from Laskar et al (2004b), Mars calculated from Laskar et al (2004a).



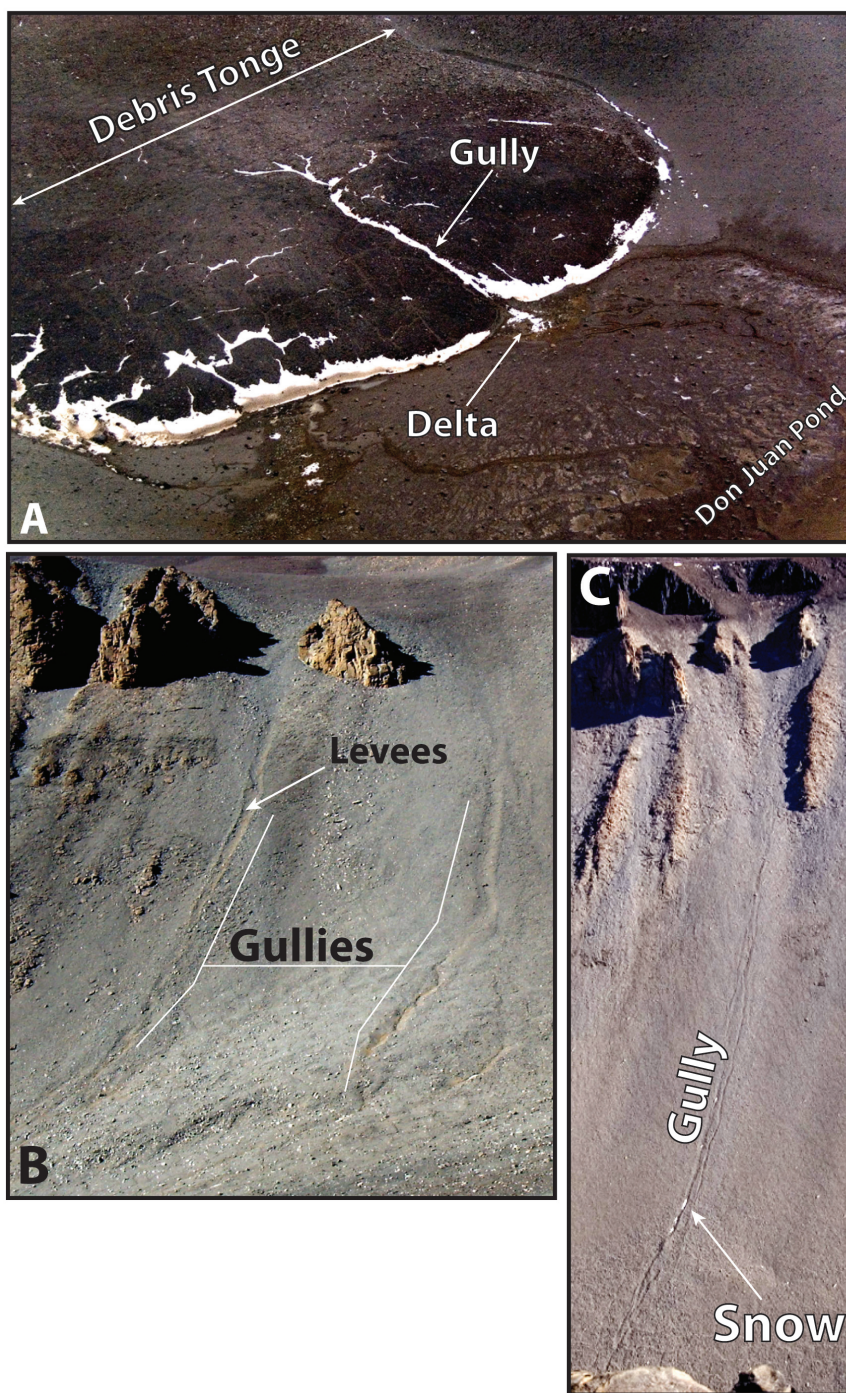
Chapter 4, Figure. 1



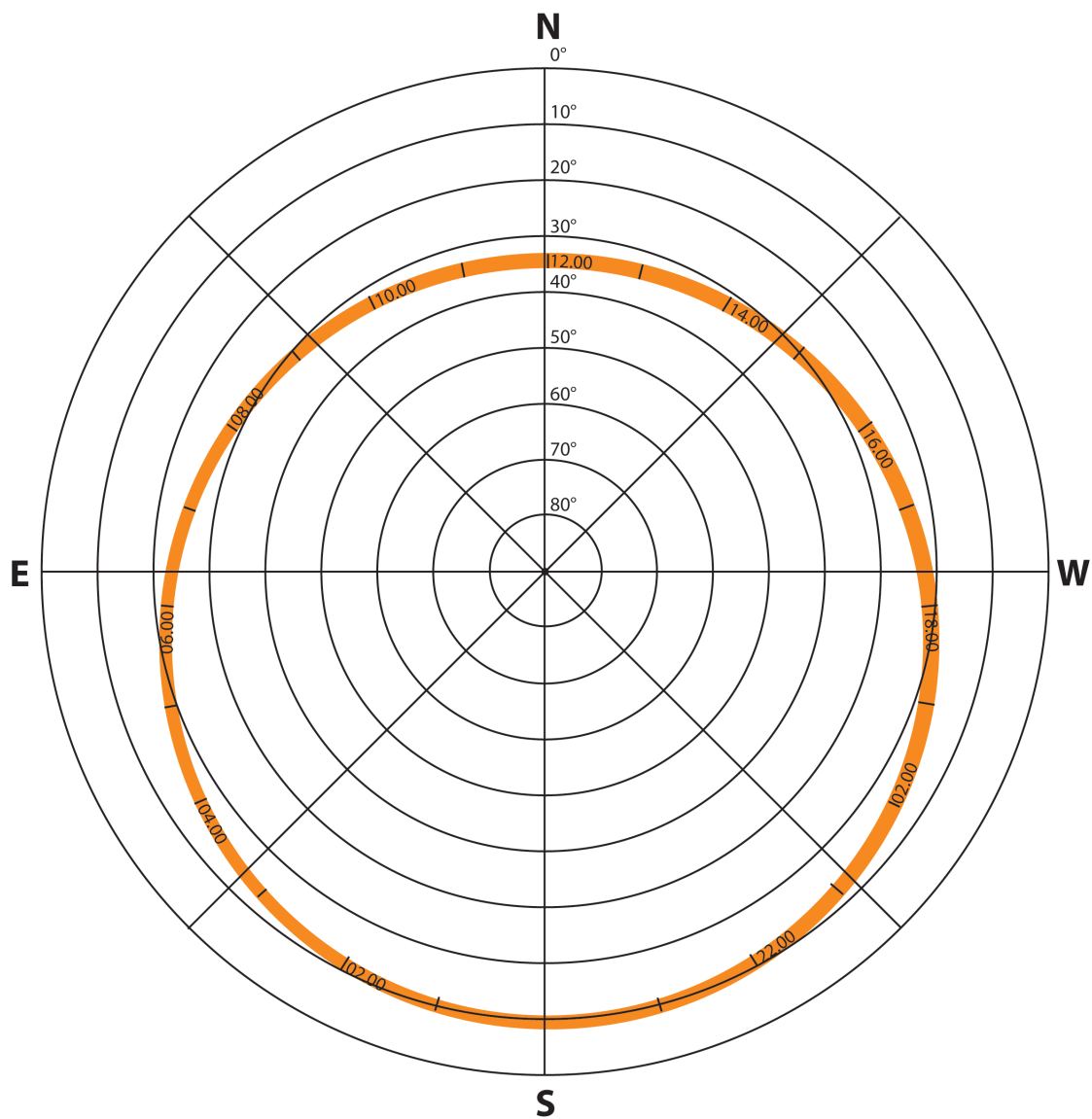
Chapter 4, Figure. 2



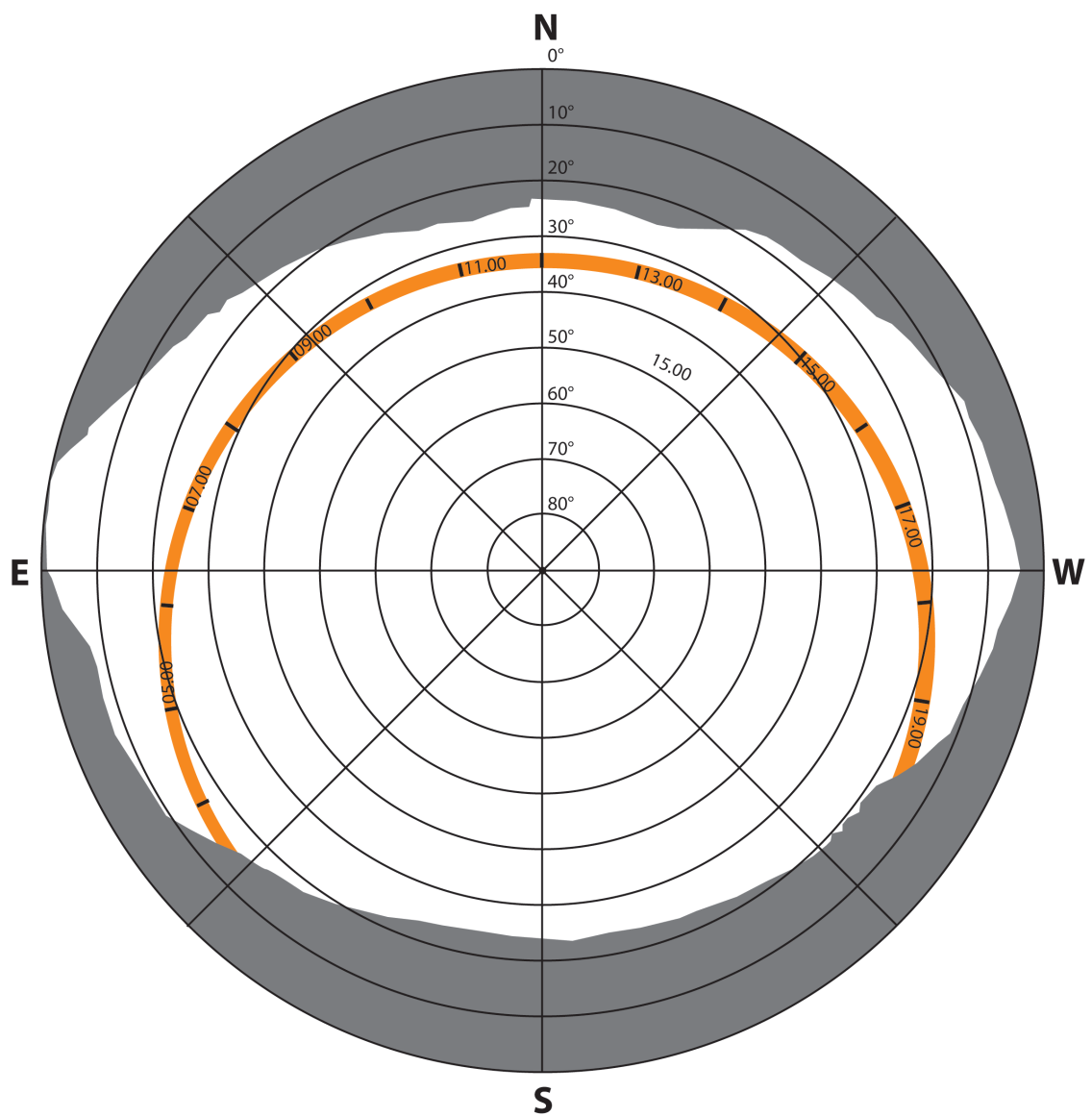
Chapter 4, Figure. 3



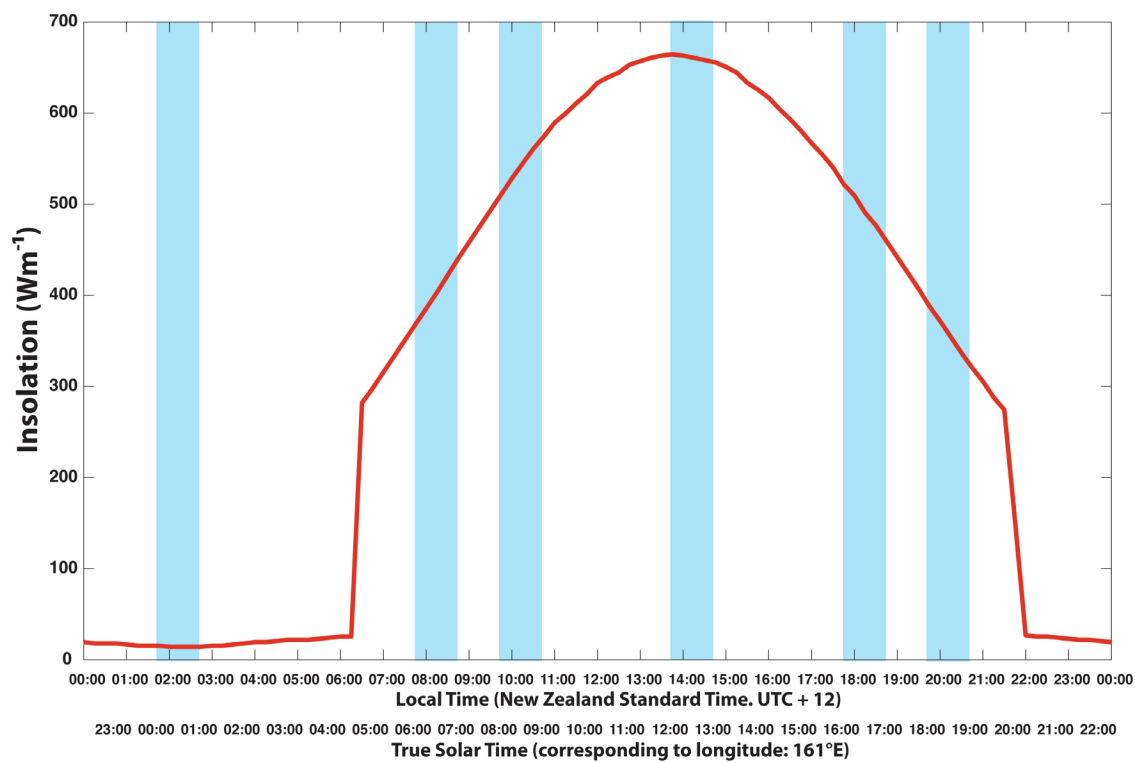
Chapter 4, Figure. 4



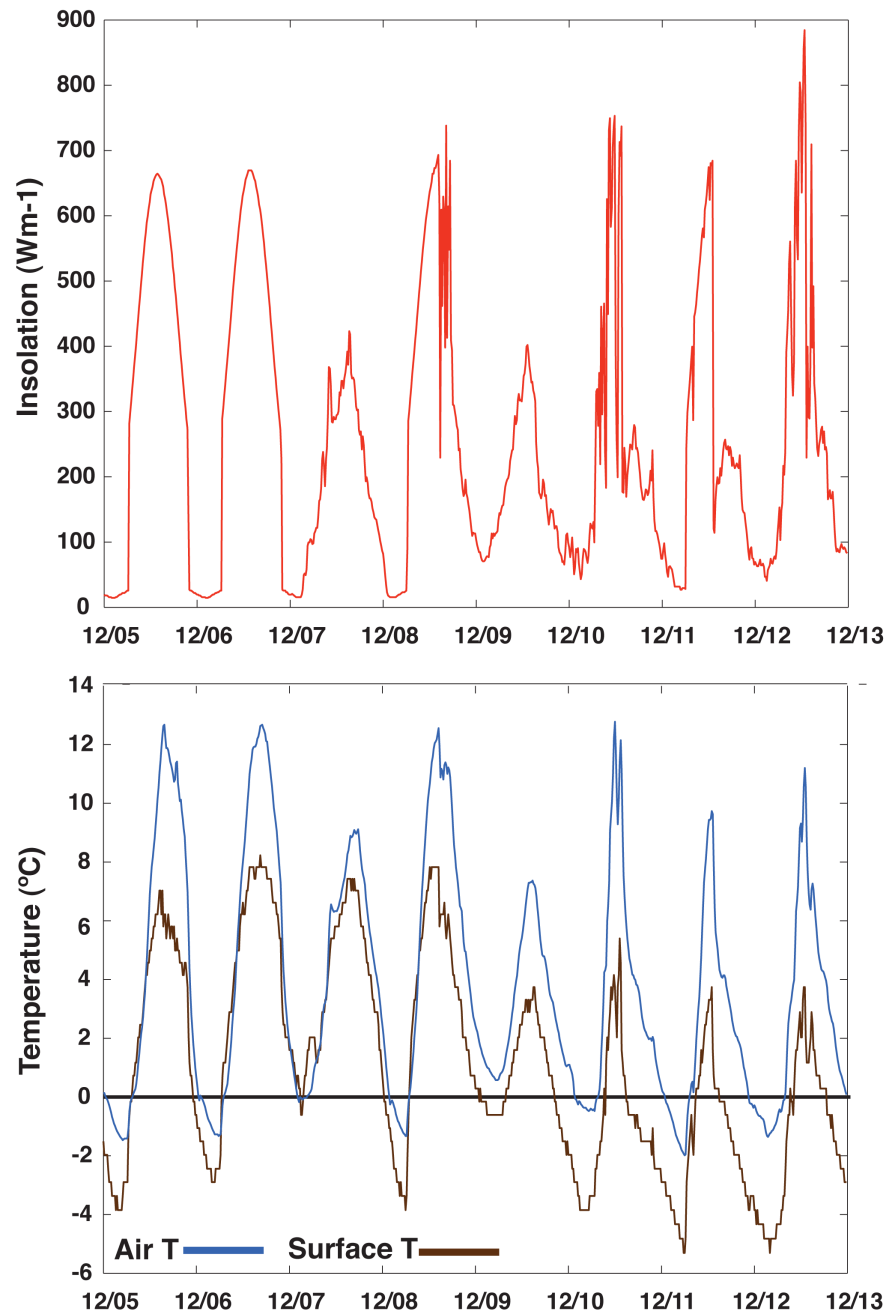
Chapter 4, Figure. 5



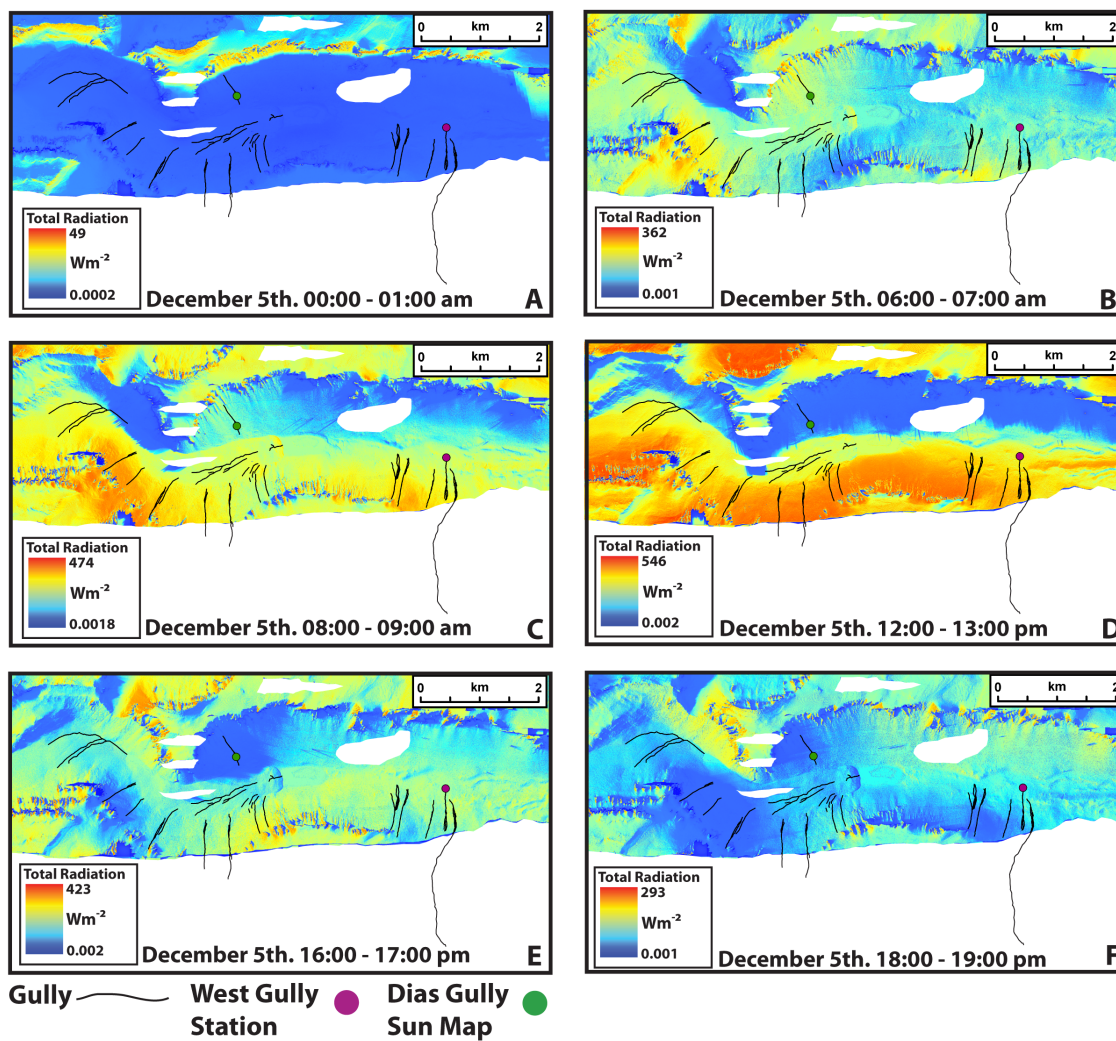
Chapter 4, Figure. 6



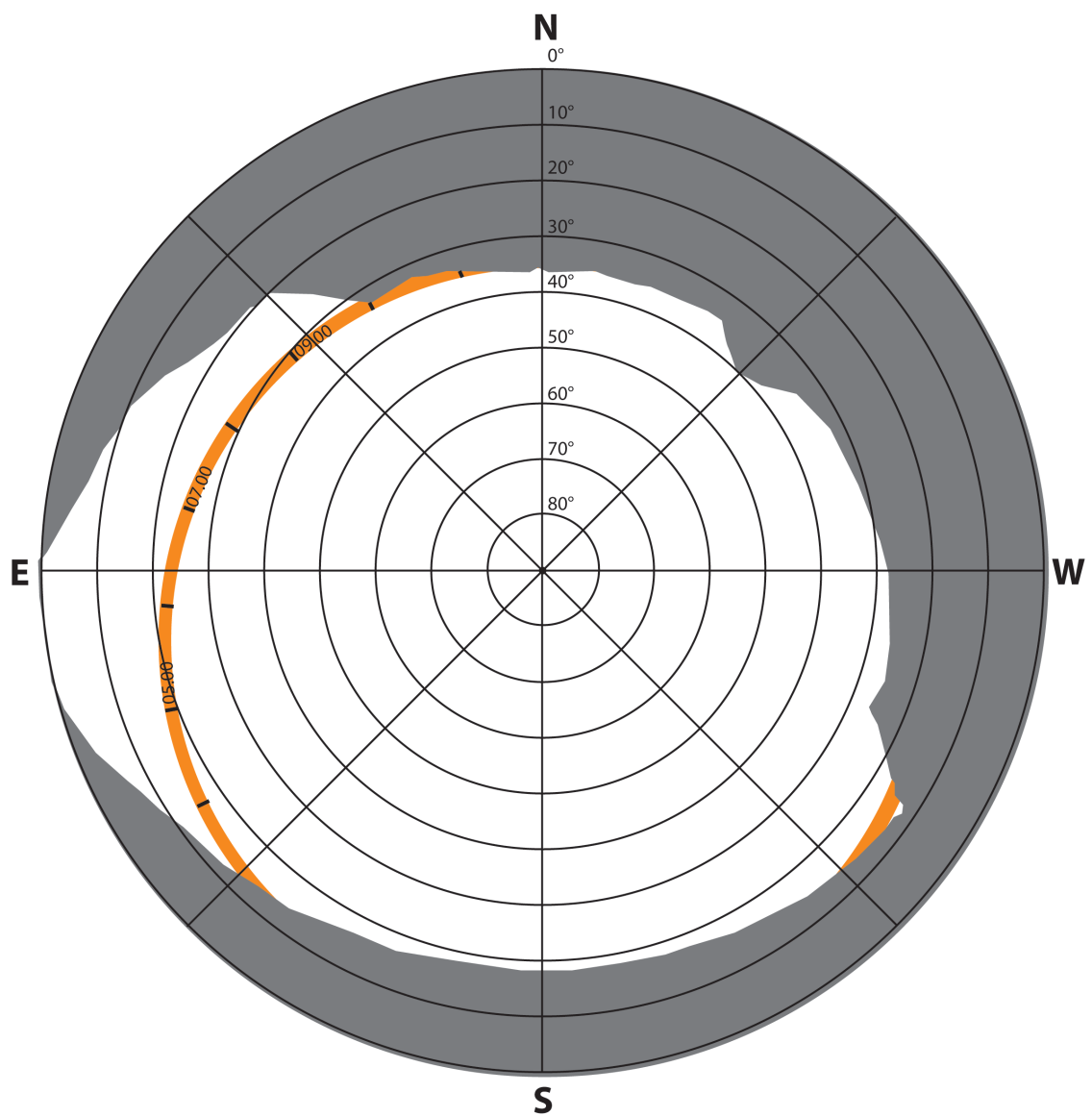
Chapter 4, Figure. 7



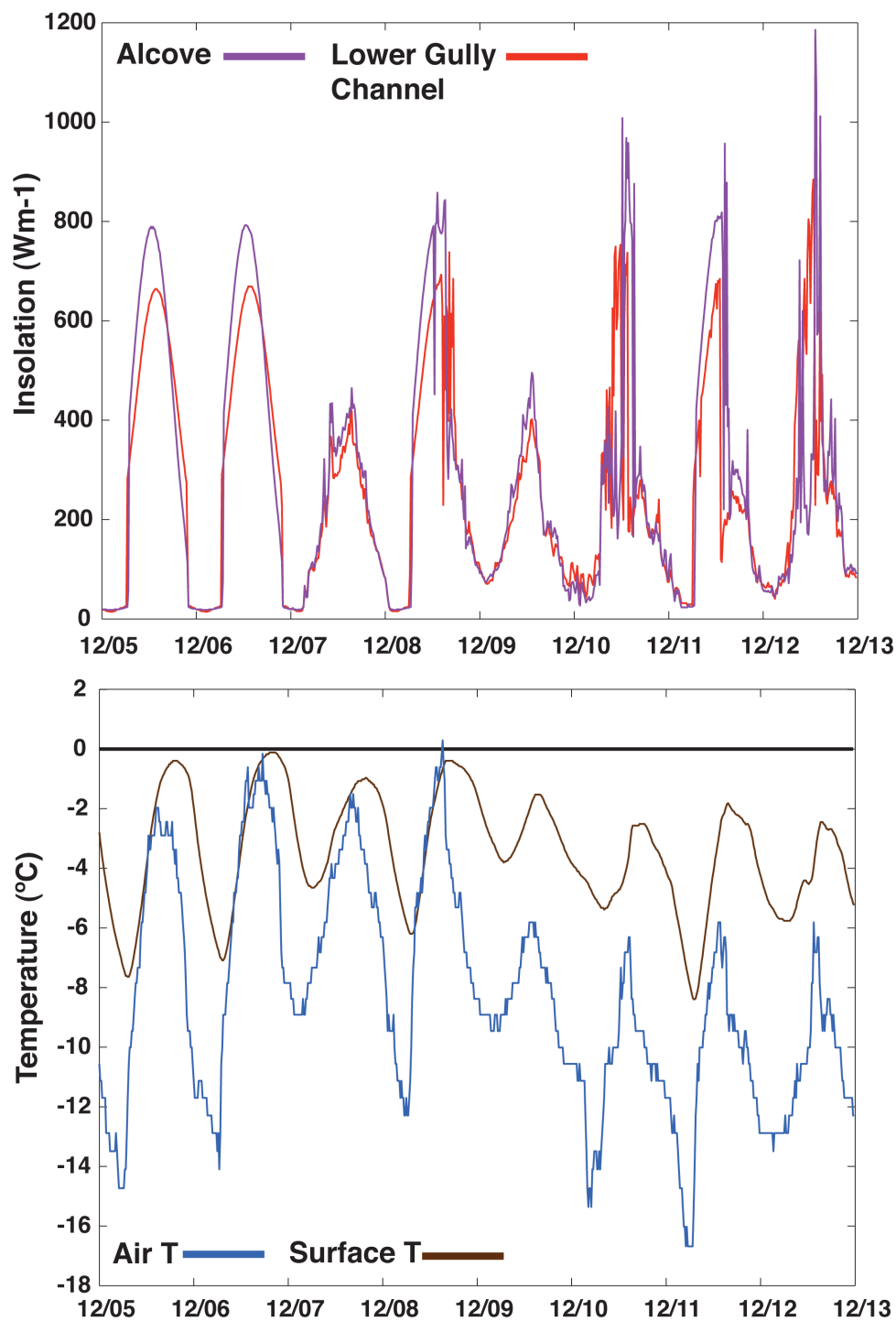
Chapter 4, Figure. 8



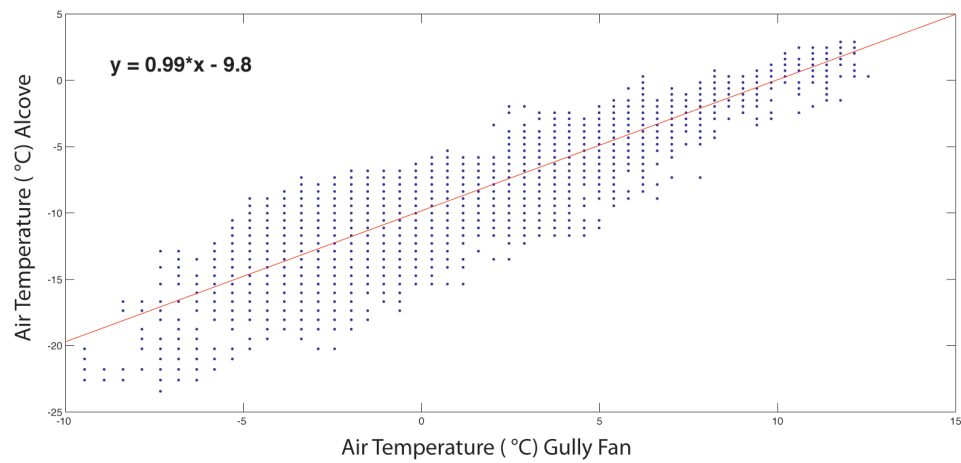
Chapter 4, Figure. 9



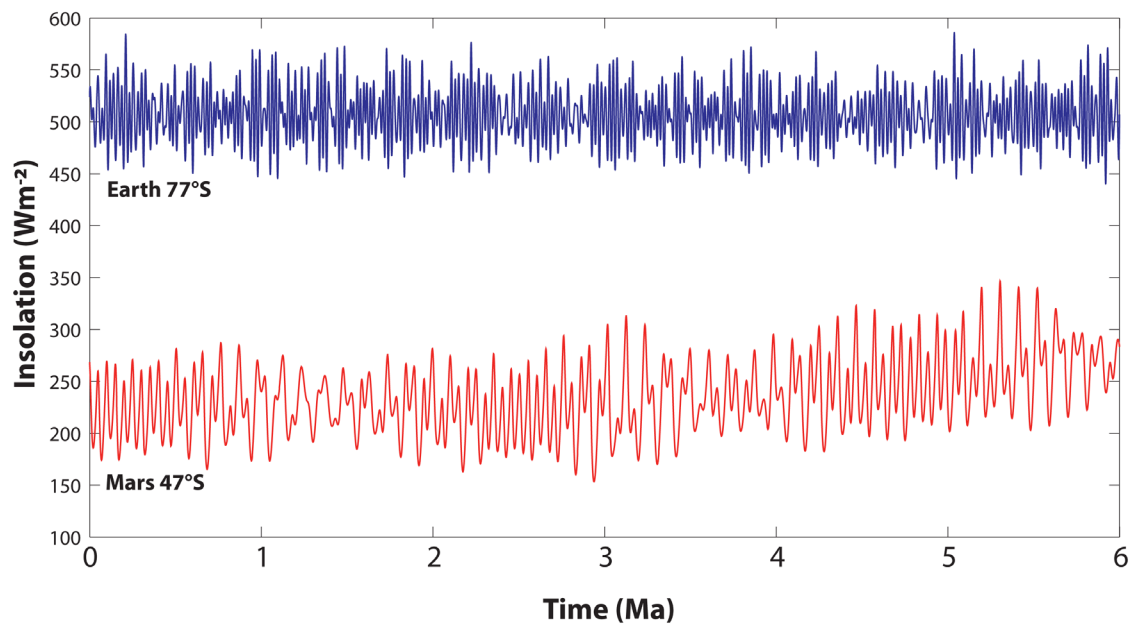
Chapter 4, Figure. 10



Chapter 4, Figure. 11



Chapter 4, Figure. 12



Chapter 4, Figure. 13

Chapter Five

Gully Formation on Mars: Two Recent Phases of Formation Suggested by Links Between Morphology, Slope Orientation and Insolation History

Gareth A. Morgan

Abstract

The unusual 80 km diameter Noachian-aged Asimov crater in Noachis Terra is characterized by extensive Noachian-Hesperian crater fill and a younger superposed annulus of valleys encircling the margins of the crater floor. These valleys provide an opportunity to study the relationships of gully geomorphology as a function of changing slope-orientation relative to solar insolation. We found that the level of development of gullies was highly correlated with slope orientation and solar insolation. The largest and most complex gully systems, with the most well-developed fluvial landforms, are restricted to pole-facing slopes. In contrast, gullies on equator-facing slopes are smaller, more poorly developed and integrated, more highly degraded, and contain more impact craters. We used a 1D version of the Laboratoire de Météorologie Dynamique GCM, and slope geometries (orientation and angle), driven by predicted spin-axis/orbital parameter history, to assess the distribution and history of surface temperatures in these valleys during recent geological history. Surface temperatures on pole-facing slopes preferential for water ice accumulation and subsequent melting are predicted to occur as recently as 0.5 – 2.1 Ma, which is consistent with age estimates of gully activity elsewhere on Mars. In contrast, the 1D models predict that water-ice cannot accumulate on equator-facing gullies until obliquities exceed 45°, suggesting they are unlikely to have been active over the last 5 Ma. The correlation of the temperature predictions and the geological evidence for age differences suggests that there were two phases of gully formation in the last few million years: an older phase in which top-down melting occurred on equator-facing slopes and a younger more robust phase on pole-facing slopes. The similarities of small-scale fluvial erosion features seen in the gullies on Mars and those observed in gullies cut by seasonal and perennial snowmelt in the Antarctic Dry Valleys supports a top-down melting origin for these gullies on Mars.

1. Introduction

The discovery of gullies on Mars attracted significant attention because of the apparent role of liquid water in carving apparently modern landscapes (Malin and Edgett, 2000a). Gullies were initially interpreted to be the result of groundwater discharge (Malin and Edgett, 2000a; Heldmann et al, 2004; 2005). Further analysis of the current metastability of liquid water on the surface of Mars generated alternative explanations, including atmospherically deposited sources of water (Costard et al, 2002; Hecht, 2002; Christensen, 2003; Dickson. et al, 2007; Head et al., 2008; Williams et al; 2009). Global surveys have demonstrated that gullies are limited to latitudes $>30^\circ$ (Malin and Edgett., 2000a) and that they have a preference for pole-facing orientations below about 45° latitude (Costard et al, 2002; Heldmann and Mellon, 2004; Dickson et al, 2007a).

Costard et al (2002) used a one-dimensional version of the atmospheric LMD/GCM (Forget et al., 1999) to demonstrate that ice accumulation and near-surface melting of ground ice could account for this spatial distribution. Their mechanism involved the accumulation of near-surface water ice on pole-facing slopes and its preservation due to the low surface temperature maintained by a seasonal CO_2 frost cover. Springtime removal of the CO_2 frost, and the subsequent rapid heating of the underlying water ice, was invoked as a means of water-ice melting, flow imitation, and gully formation.

Recent high-resolution studies of individual gully sites have revealed a wide range in gully morphology, including the occurrence of fine scale channel forms such as streamlined islands, braiding and terraces (McEwen et al., 2007), all indicative of fluvial erosion. Evidence for different forms of fluvial erosion (debris flows, stream flow incision, etc.) suggest that varying ratios of water and sediment have eroded gullies on Mars. Analysis of gullies in the Mars-like Antarctic Dry Valleys (ADV) (Marchant and Head, 2007) has shown that snowmelt from seasonal and perennial snow deposits can serve as a source for seasonal gully activity in a

hyper-arid polar desert setting (Morgan et al., 2008; Dickson et al., 2007b; Levy et al., 2007). In the ADV, precipitation occurs only as snow, with annual precipitation of only a few centimeters. This snow is redistributed by the wind and concentrated during the winter into seasonal and perennial snowpacks within topographic hollows (alcoves and gully channels). Gully activity is initiated by the melting of these snowpacks in the spring by direct solar insolation, forming localized surface runoff and limited fluvial activity. Recent modeling by Mischna et al. (2003) has shown that snowfall on Mars is possible during periods of higher obliquity in the regions where gullies are located. Conditions favorable for the accumulation and preservation of snow and ice must be maintained long enough, prior to the onset of melting, in order for liquid water to be produced on Mars to carve the gullies (Hecht, 2002). The fluvial characteristics of martian gullies (McEwen et al., 2007), their distinctive latitude dependence (Malin and Edgett, 2000a), their close association with recent ice-rich mantles (Milliken et al., 2003, Head et al., 2003; Christensen, 2003) and lobate glacial-like lobes (Head et al., 2008), and evidence for snow and ice accumulation in current gully alcoves and channels (Head et al., 2008), all point to the potential role of top-down melting of accumulations of snow and ice as the source of water forming the gullies.

Recent Mars studies where gullies are found on slopes of all orientations have noted aspect-dependent variations in morphology related to slope orientation (Reiss et al., 2008). We test the viability of the top-down snow-melt model as an explanation for gully activity on Mars by investigating whether variations in insolation conditions favorable for the annual accumulation and subsequent melting of snow can account for aspect dependant morphological differences observed in gullies. We apply the Costard et al. (2002) 1D version of the LMD GCM (used to explain the global spatial distribution of gullies) to the study area in Noachis Terra. We use site-specific topographic orientation and slope parameters as inputs for the model to assess the current distribution of temperatures. We then examined the history of insolation and surface temperatures by using the predicted spin-axis/orbital history for the last 20 million years derived

by Laskar et al. (2004). We then assessed whether predicted variations in insolation-related surface temperatures could account for observed differences in gully morphology and ages of gully activity.

2. Location and Morphology of the Gullies

The study site chosen for our investigation is within the, 80 km diameter, Noachian-aged degraded Asimov crater located at 46° S, 5°E, within Noachis Terra (Fig. 1). The floor of the crater is unusual in that it contains an annulus of deep valleys (~2 km maximum depth) and an irregular central depression. Studies of Asimov crater and similar examples suggest that the initially fresh crater was infilled with material and that the valleys were subsequently formed by the preferential removal of the crater fill along the interior walls of the crater (Schultz and Glicken., 1979; Malin and Edgett., 2000b). Gullies are located along every slope within the interior valleys and central depression, though their morphology varies significantly with slope orientation. Despite morphological variations, the gullies are all composed of the three basic morphologic units used to define martian gullies (alcove, channel and fan; Malin and Edgett., 2000a) and therefore, are considered to represent morphologic variations of a single landform type.

A thick, resistant, cliff-forming rock unit displaying columnar jointing and interpreted to be a lava flow (Morgan et al., 2009) is present along the upper portions of the valley walls and provides a source of boulders to the slopes below. Gullies typically originate close to this layer, suggesting that there is a relationship between the two. This type of relationship has been interpreted as evidence for a groundwater source through the containment of a perched aquifer by the rock layer (Gulick et al., 2007). However, the occurrence of gullies along isolated ridges formed by the narrow divides between adjacent valley systems is inconsistent with the groundwater hypothesis for gully formation (Malin and Edgett, 2000a; Heldmann et al, 2004;

2005) and instead supports gully activity resulting from an external water source. It also implies that variations in gully morphology are the result of external forcing and is not influenced by endogenic processes undetectable from the spacecraft data.

The largest and most complex gully forms are located on pole facing (PF) slopes, and typically consist of multiple branching tributaries that originate at the base of the rock layer and merge together downslope to form single channels (~80 m wide) that open up into depositional fans (Fig. 2). Within the gully channels, smaller scale ~10 m wide internal channels are present, which at HiRISE resolution (sub-meter) can be seen to contain fine-scale fluvial-like features, including terraces and braided channels (Fig. 2). These features are similar to those observed to form in the active gullies in the ADV and suggest that fluvial erosion was involved in the formation of the gullies (e.g. Morgan et al, 2008). Abandoned channels and distinctive fan stratigraphy (Fig. 2a) suggest that multiple episodes of activity have been recorded in the gully morphology.

In contrast to the PF gullies, the gullies on equator facing (EF) slopes have smaller, thinner (~20 m wide) channels that branch out from the apex of well-defined 200 - 500 m wide, cusped alcoves that are cut into the exposed rock layer at the summit of the slopes (Fig. 2b). EF slopes are steeper than PF slopes, suggesting that the latter have experienced more erosion. Large amounts of debris (including 10 m diameter boulders) are visible within the alcoves and spread out down slope forming talus cones into which the gully channels are eroded. This suggests that gully activity along the EF slopes has been accompanied by dry mass-wasting processes in the form of rockfalls and debris slides. Together, these characteristics argue for the involvement of lower volumes of water in the erosion of the EF gullies relative to that involved in gullies on the PF slopes. Similar slope asymmetry is found within the ADV, with gentler slopes resulting from a greater availability of liquid water (Marchant and Head, 2007). The east and west-facing gullies are morphologically similar to the PF gullies in terms of the level of incision, although they do not exhibit the same degree of complexity or finer-scale bedforms.

Pole-facing gullies appear fresher and more well-developed than equator-facing gullies (Fig. 2), suggesting that the PF gullies have been active more recently than the EF gullies. Determining an absolute age for the two gully systems using crater size-frequency distribution data is difficult due to the small sample areas involved; furthermore, the steep angles on which the gullies have formed ($>15^\circ$) makes them prone to failure and thus degradation resulting in the loss of impact craters. Nevertheless, a survey of all available MOC (22) and HiRISE (8) images over the entire study region revealed that ~ 70 craters with a diameter $> 5\text{m}$ (including 6 craters with a diameter of $>100\text{ m}$) are present on EF gullies relative to only ~ 15 on PF gullies. This suggests that the EF facing gullies have not been active as recently as the PF which is consistent with the morphological interpretation.

What are the causes of orientation-dependant differences in gully morphology? Examination of the circular valley systems shows that the aspect-dependence of gully morphology was maintained regardless of whether the gullies were eroded into the crater fill material or the opposite valley side, along the interior of the crater walls. This suggests that insolation, not slope composition, was the critical factor in the development of gullies of different morphology.

3. Model Results.

In order to investigate the differential insolation conditions related to aspect we employed a one-dimensional version of the GCM model developed by Costard et al (2002). Within the model the diurnal and seasonal surface temperatures are derived from the balance between the radiative and turbulent fluxes, thermal conduction into the regolith and CO_2 condensation and sublimation. We used the 1D model to predict the maximum annual surface temperatures at the study site under three different scenarios: 1) present insolation conditions, 2) maximum insolation conditions during the period of 35° average obliquity values (868 Ka), and 3) maximum insolation conditions during the period of 45° average obliquity values (9.11 Ma). We assessed

the results in terms of conditions that might permit the accumulation, and subsequent melting, of snow and ice (Fig. 3 and 4). Our simulations demonstrate that for the majority of the year the PF slopes temperature at all obliquities is constrained by the frost point of CO₂ (as was highlighted by Costard et al, 2002) and thus remains at ~150 K (Fig. 3). As this temperature is below the frost point of H₂O, the PF slopes will also be favorable environments for the accumulation of snow and ice, provided that there is a source of snow. Under current conditions snowfall is less likely to occur, although seasonal ice accumulations have been reported in this latitude range (Schroghofer and Edgett, 2006; Head et al., 2008). During periods of higher obliquity, greater amounts of snow and ice may have been deposited as a result of the increased sublimation of the residual summer cap (e.g., Mischna et al., 2003). Snow and ice deposits could have been built up on the PF slopes through the winter by the redistribution of snow by the winds in a similar manner to that observed in gullies in the ADV (e.g. Morgan et al., 2008). During the spring, the rapid removal of the CO₂ frost cover permits enhanced heating of the surface. The exposure of any H₂O snow deposits accumulated during the winter would prevent this temperature increase from immediately reaching the H₂O melt point due to the relatively high albedo of snow (0.39, for slightly dusty snow: Williams et al., 2008). However, modeling by Williams et al (2009) has demonstrated that due to the expected dust content of snow on Mars, the initial effects of sublimation would lead to the generation of a surface lag and thus reduce the albedo sufficiently (0.13, albedo of dust layer; Williams et al., 2008) to permit melting to occur (Fig. 3). In addition to this, localized dust storms generated by the high thermal gradient between the receding CO₂ frost cover and the exposed ground could deposit a thin layer of dust over exposed snowpacks. Williams et al. (2009) has shown that the melting of small dusty snowpacks (~ 1 cm thick) can generate 1 mm of runoff/m². This volume of snowcover over an average PF alcove (~2 km²) would generate ~2000 m³ of runoff, enough to fill a 200 m long section of inner channel (10 m x 1 m; Fig 2c) at bankfull conditions (assuming no losses to infiltration, evaporation and freezing). This level of activity would require repeated cycles of accumulation and melting to generate the ~

2 km channels and is consistent with geomorphic evidence for multiple episodes of activity. Our results suggest that gully activity along PF slopes would have been favored beginning in the recent past during phases of peak obliquity ($\geq 35^\circ$) in the period from ~ 0.5 -2.1 Ma (Fig. 4). These periods are consistent with age estimates of gully activity elsewhere on Mars (Reiss et al., 2004; Schon et al., 2009).

The model results for the equator-facing slopes (Fig. 3b) are significantly different than the pole-facing slope results. On PF slopes, gully activity is only likely to have taken place during periods when obliquities are $>45^\circ$, despite conditions favorable for melting during all obliquities. For obliquities below 45° the winter (Ls $90^\circ - 180^\circ$) surface temperatures do not become sufficiently cold for CO₂ frost to be deposited (Fig. 3b). As a result of this temperatures remain largely above the frost point of H₂O, preventing the accumulation of ice. In addition to this the surface temperatures warm gradually to the melt point of H₂O during the spring and thus are unfavorable to the formation of meltwater (Hecht., 2002, Costard et al., 2002). At obliquities $>45^\circ$ the annual temperature regime is similar to the PF slopes, and CO₂ condensation can occur (Fig. 3a). However, the potential accumulation period (Ls $70^\circ - 190^\circ$) is close to half the length that is experienced during the same period on the PF slope (Ls $20^\circ - 250^\circ$), and as a consequence the gullies may have had less time to accumulate snow (and thus have a lower potential supply of meltwater). This argues that the EF gullies were only active during this earlier period, prior to 5 Ma (Fig. 4). This is consistent with the higher number of superimposed craters and a more degraded appearance of EF gullies relative to PF gullies. Even during 5-10 Ma, obliquities of 45° were only achieved for 20 Kyr during each ~ 110 kyr obliquity cycle relative to the 70 Kyr spent at obliquities $> 35^\circ$ during the same time period. Hence the limited accumulation period and lower frequency of activity may account for the vast difference in gully morphology between the EF and PF slopes. Due to the influence of eccentricity on insolation, not all of the periods of high obliquity will correspond to the same amount of insolation (Fig. 4). Therefore, gully activity is

expected to vary in response to this over the last 20 Ma.

Figure 5a shows the spatial distribution of the PF and EF gullies that display the most prominent differences in geomorphological characteristics described above. The two end member gully types were mapped based on the following morphological characteristics: PF end member gullies were defined by broad alcoves that contain multiple tributaries that coalesce to form single large sinuous channels. EF gullies were defined by linear channels that originate from the base of well-defined cusped alcoves that are cut into the rock unit. The difference in valley depth (Fig. 1) appears to have had a limiting effect on both gully types. Hence, the two end members were not defined by specific scale ranges as both groups were affected by the lengths of the slopes they were on. The gullies on slopes not highlighted on figure 5a consisted of morphologic characteristics representative of both end member types. Figure 5 b - d shows the model results based on the aspect and slope values for the study site derived from a 200 m/pixel HRSC Digital Elevation Model. The value represented on the maps is the time taken for the temperature to increase from 150 K to the melting point. Hence, the smaller the value the higher the potential for gully activity to have occurred (assuming that a dust layer is produced over the snowpacks to reduce the albedo sufficiently). Larger time periods will increase the effect of loss through sublimation, reducing the volume of the snowpack, possibly to the point at which it is completely removed before melting could occur. Regions of no activity (such as EF slopes at obliquities $< 45^\circ$) correspond to conditions where either the temperature never becomes cold enough for CO₂ frost to form or the melt point is never achieved. The model results show good agreement with the map of gully morphology (Fig. 5a). In addition to representing the spatial distribution of the extreme insolation environments experienced by EF and PF slopes, figure 5 also provides an insight into gully formation at other orientations. Gullies on east and west-facing slopes experience gully activity index values between those of the PF and EF slopes, which is consistent with the gullies on these slopes consisting of morphologies between the two end members.

Further high resolution modeling is required to fully establish the legitimacy of the wind as means of concentrating precipitation into snowpacks and full account has to be taken for the physics responsible in melting snow in order for the application of the ADV snow melt model to Mars to be fully tested. Nevertheless, from a first order approach, the top-down melting of gully formation model presented here is consistent with the morphological variations displayed in the study area and might provide an insight into morphologic variations observed elsewhere on Mars.

Acknowledgements

Special thanks are extended to my advisor **James W. Head** and to **Francois Forget**, **Jean-Baptiste Madeleine**, and **Aymeric Spiga** for extensive discussions on this work during the formulation of the ideas and content of this Paper. I look forward to additional assistance and collaboration with all of them as this contribution is prepared for publication.

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Figure Captions.

Fig. 1. Study region: Asimov crater in Noachis Terra. The boxes represent the location of the gullies in Fig 2. HRSC orbit 1932_0000 (Image and DTM).

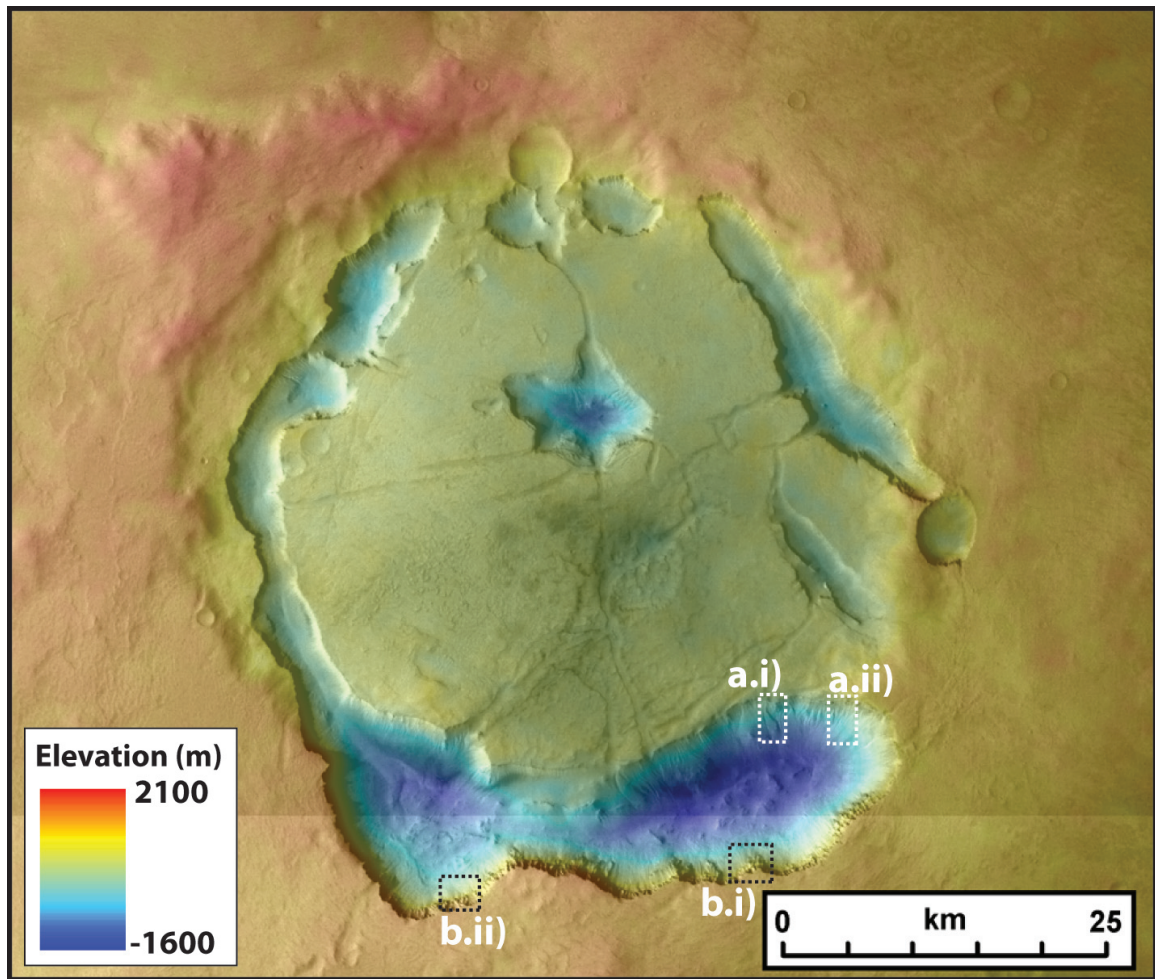
Fig. 2. High-resolution images of gullies with different orientations. (ai and ii) Pole-facing gullies; these are the most complex and incised gullies found in the study area. (aiii) Close-up of box in (aii) showing fine-scaled channel features suggestive of fluvial erosion. (bi and ii) Equator-facing gullies; these represent the simplest of the gully types. (iii) Close-up of box in (bii) showing linear channels superimposed by impact craters, suggesting that the PF gullies have been active more recently than the EF gullies. (ai) MOC: E0301360 (aii and aiii) HiRISE: PSP_004091_1325. (bi) MOC: E1101724 (bii and biii) HiRISE: PSP_00 6926_1320.

Fig. 3. Maximum surface temperatures (K) over a year at 46 °S for a 25° polar facing slope (a) and a 25° equator-facing slope (b) under present conditions, and under the conditions

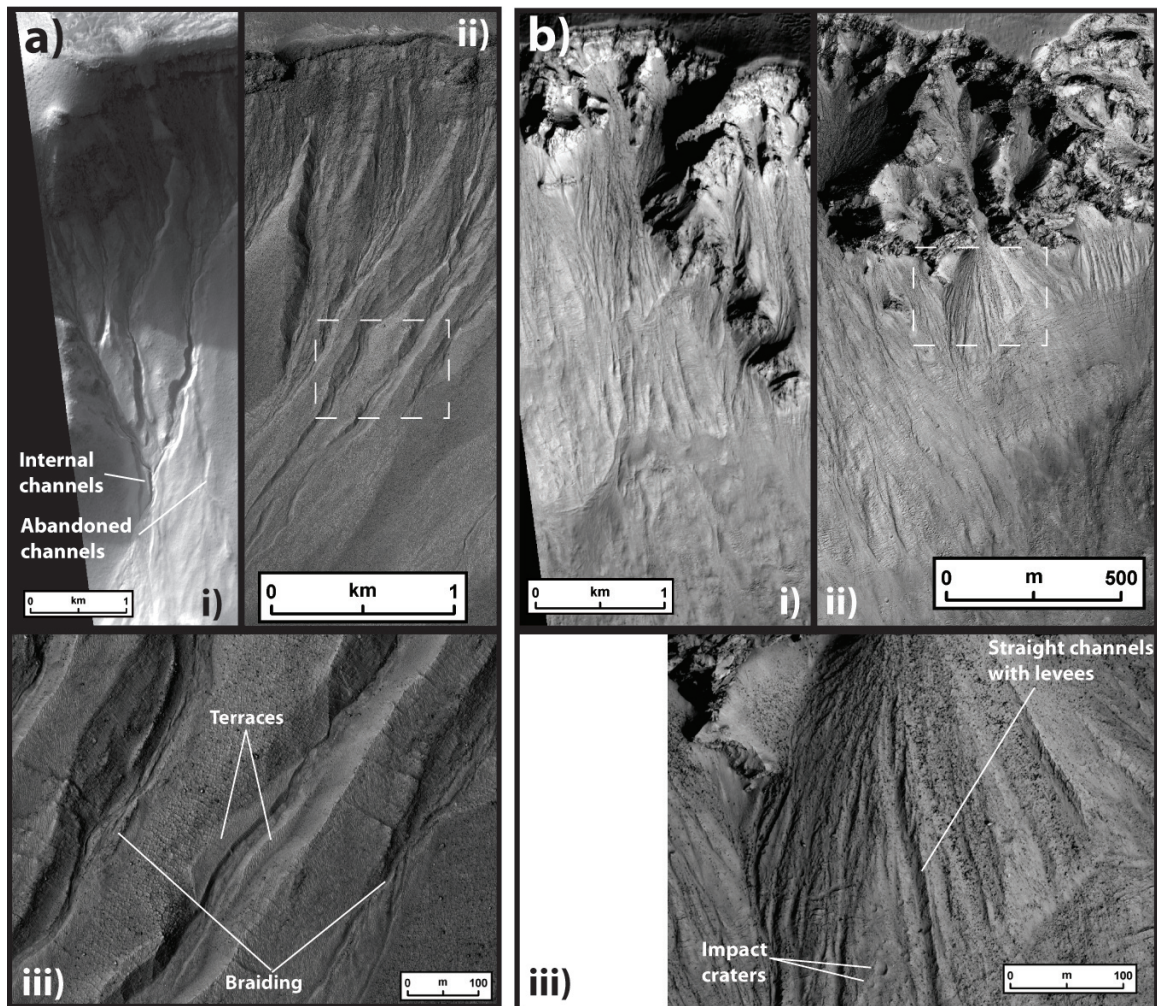
dictated by the orbital parameters (obliquity, eccentricity and longitude of perihelion) predicted by Laskar et al (2004) for 868 Ka (obliquity 34.9°) and for 9.1 Ma (obliquity 46.3°).

Fig. 4. Obliquity and insolation values for 46°S simulated by Laskar et al (2004). The potential periods of the most recent gully activity for PF and EF slopes are plotted based on the results of the model and the morphologic investigation.

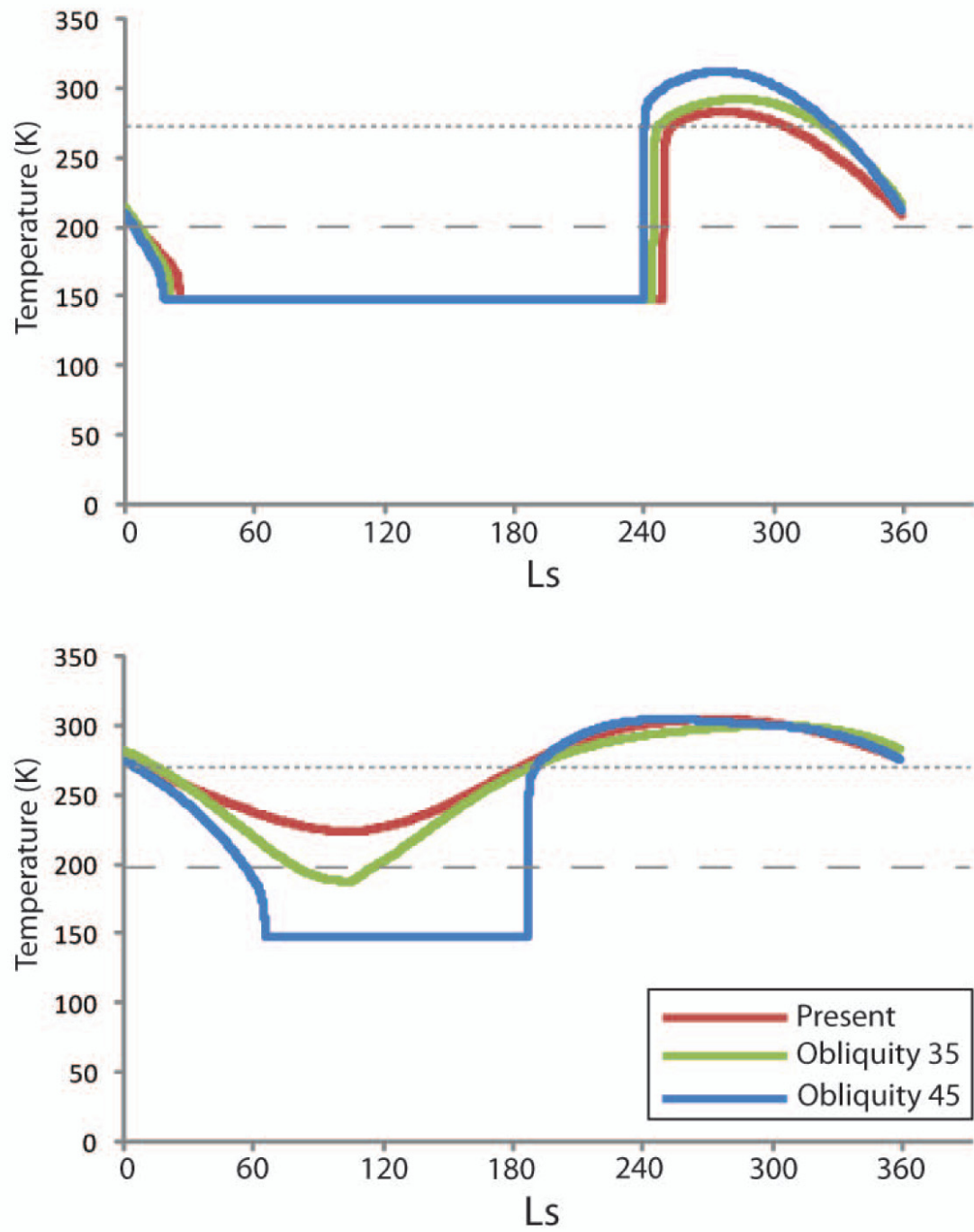
Fig. 5. (a) The location of the two morphologic gully end member types. (b – d) The spatial representation of the model results based on the aspect and slope values for the study site derived from a 200 m/pixel HRSC DTM. The value represented on the maps is the length of time (in sols) taken for the spring maximum surface temperatures to rise from 150 K (frost point of CO₂) to the melting point of water. No activity corresponds to conditions where either the temperature never becomes cold enough for CO₂ frost to form or the melt point is never achieved. Hence, the lower the number the larger the potential for gully activity to occur in the year from a combination of snow deposits being stable long enough to accumulate into sufficient volumes and then being able to melt rapidly. b) The present. c) 868 Ka (Obliquity 34.9°). d) 9.1 Ma (Obliquity 46.3°).



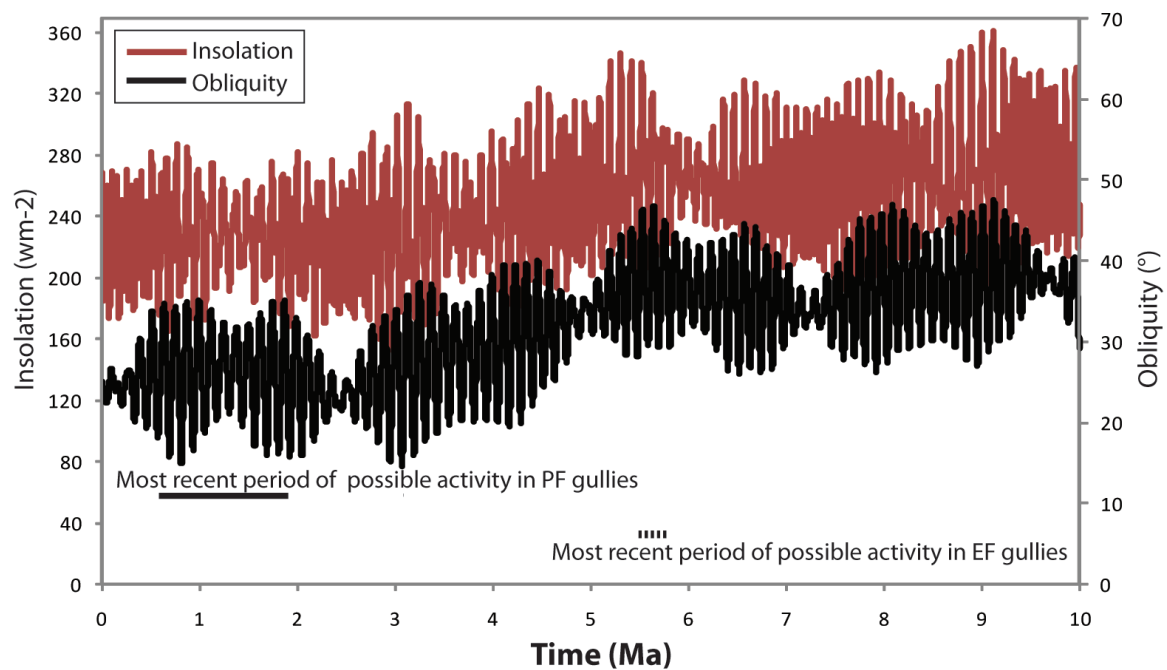
Chapter 5, Figure. 1



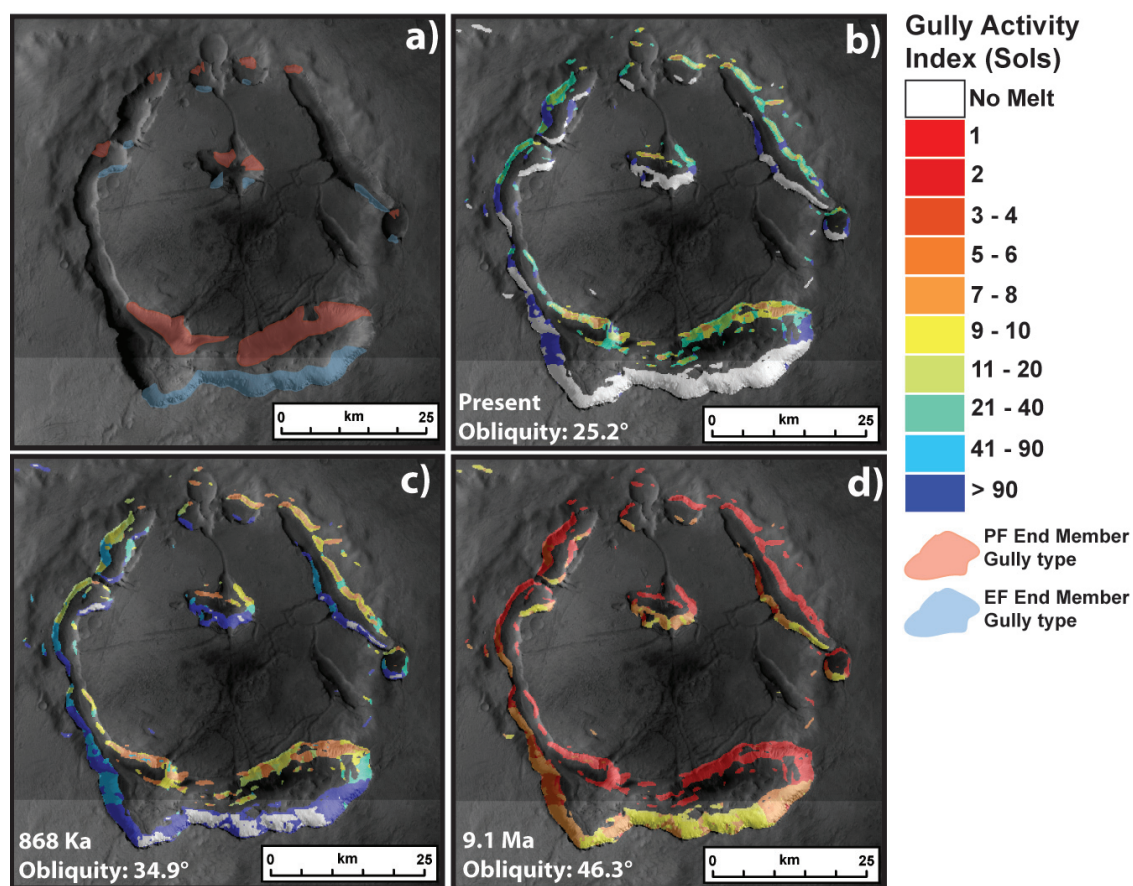
Chapter 5, Figure. 2



Chapter 5, Figure. 3



Chapter 5, Figure. 4



Chapter 5, Figure. 5

Chapter Six

Age Relationships of Lobate Debris Tongues and Gullies in a Unique Crater Environment in Noachis Terra, Mars: Comparison to Mars-like Environments in the Antarctic Dry Valleys

Gareth A. Morgan

Abstract

The Amazonian period of the history of Mars has largely been described as a ‘cold and dry’ climate characterized by a hydrological cycle dominated by water in the solid and gaseous phase, recycled through the poles and regolith of the high latitudes. Recent analysis of high-resolution image data have revealed an array of young landforms associated with both ice and liquid water at equatorial and mid-latitudes, where it is currently unstable, suggesting significant climatic change during the Amazonian. Here we report on a detailed investigation into valley systems within Asimov crater, a degraded crater in Noachis Terra that have produced a unique environment to study the geomorphic signals of Amazonian climate change. The occurrence of steep slopes ($>20^\circ$), relatively narrow (hence sheltered) valleys and a source of debris have provided favorable conditions for the preservation of ice and the formation of gullies. To aid our interpretation of the landform assemblages, we conducted field investigations in the McMurdo Dry Valleys of Antarctica (ADV). Detailed mapping in the ADV reveals terrestrial analogs for the origin of lobate debris tongues (LDT) in Noachis Terra (i.e., upper Beacon Valley), and subsequent modification of LDT and gully incision (i.e., present-day South Fork, Wright Valley). The morphologic change is best explained as a shift in martian climate, from one compatible with excess snowfall, and formation of debris-covered glaciers, to one incompatible with excess snowfall and glaciation, but consistent with minor snowmelt and gully formation. Available dating suggests that the climate transition occurred > 10 Ma, prior to the formation of other small scale ice rich flow features identified elsewhere on Mars that have been interpreted to have formed during the most recent phases of high obliquity. This suggests that multiple climatic shifts have occurred over the last tens of millions of years of martian history.

1. Introduction

The presence of extensive fluvial features (e.g., widespread dendritic valley networks) on ancient Martian terrain suggests that a relatively ‘warm and wet’ climate was prevalent early in the planet’s history (Carr, 1996). Such conditions provide a strong contrast to the hyper-arid, extremely cold climate (characterized by temperatures and pressures largely below the triple point of H₂O) that is thought to have persisted throughout the Amazonian, up to the present (e.g., Bibring et al., 2006). Within the context of a general Amazonian polar desert-like environment (Baker, 2001), the hydrological system is horizontally layered and most water resides in a global cryosphere and the polar caps, with minute amounts in the atmosphere. In contrast to the Noachian, there is little evidence of liquid water substantially shaping the surface. Marchant and Head (2007) have emphasized, however, that microenvironments can exist within a global and regional context in which conditions can change significantly due to local insolation conditions. They show how in the Mars-like Antarctic Dry Valleys and on Mars as a function of latitude and local microenvironment, conditions can be different than inferred at the regional and global scales, and can even include melting and fluvial activity, despite the absence of pluvial activity. Furthermore, Laskar et al. (2004) have documented the extreme variations in insolation conditions that are caused by the substantial range of spin-axis/orbital parameter variations of Mars. These results show that even in the hyper-arid cold desert environment of the Amazonian of Mars, ice can be transported from polar reservoirs to the mid-latitudes (Head et al., 2006a,b) and to the equatorial regions (Head and Marchant, 2003; Shean et al., 2005, 2007a; Forget et al., 2006; Kadish et al., 2008). The placement of high-resolution instruments into Mars orbit has made it possible to observe the effects of such recent climate change and candidate melting conditions. For example, the presence of geologically young (< 1 Myr), small-scale (~1 km in length) gullies (Malin and Edgett, 2000a), argues for the localized temporal stability of liquid water at the surface within the most recent period of the Amazonian. Therefore, understanding the

detailed nature of Amazonian climate history is an important goal and is essential for assessing the potential for life to have existed on the planet over its recent history.

The thin atmosphere and low temperatures that are thought to have persisted on Mars throughout the Amazonian (Carr, 1996) mean that water-related landforms (regardless of whether in the solid or liquid phase) can be important indicators of the presence of specific environmental conditions. For example, the presence of fluvial landforms, however limited in distribution and scale, can indicate the presence of climatic conditions conducive to surface melting (i.e. surface temperatures above 273 K, assuming the absence of salts, which can enable melting point depression). The presence of deposits indicative of glaciation, without evidence of melting, signals the presence of climatic conditions permitting sufficient accumulation of snow and ice to cause cold-based glacial flow (i.e. glaciers that are frozen to their bed and flow through internal deformation rather than basal sliding). The combination of landforms indicative of synchronous glacial flow and melting might signal climatic conditions favoring wet-based glaciation. Documentation of such landforms (glacial lobes and gullies) that are *superposed* on one another would point to glacial conditions *followed* by melting conditions (e.g., Millikin et al., 2003; Head et al., 2008).

In summary, many landforms produced by water-related processes may form in a narrow range of conditions and thus can be extremely sensitive indicators of these conditions. In the terrestrial literature landform assemblages derived from a set of processes determined by the particular climatic parameters (i.e. precipitation levels and temperatures) of the climatic zones in which they inhabit have been termed *equilibrium landforms*. The mapping of such landforms have been used to classify different morphogenic regions on the Earth (e.g. Wilson, 1969; Baker, 2001). The application of such climate geomorphology is highly dependent on both the spatial scales of the landscapes being studied and temporal scales of the climate under which the landforms were generated. Hence, the term equilibrium can be viewed either as Gilbert proposed as the condition about which a landscape will fluctuate under steady state in the shorter term or

alternatively as Davis proposed as the condition the long term evolution/decay of a landscape will achieve (see Chorley et al., 1985 for more discussion). Large-scale landforms that have evolved over long periods of time and have experienced multiple climatic transitions may preserve properties inherited from previous climates. Topographical signatures of both fluvial and glacial erosion within the Antarctic Dry Valleys are a terrestrial example (Jamieson and Sugden, 2008), as is the fretted valleys and associated alcoves along the northern dichotomy boundary of Mars. These valleys show evidence of possible fluvial (Carr, 2001) and tectonic origins (McGill and Dimitriou, 1990) but appear to have been modified by processes that are largely poorly understood (Carr, 2001), but may have resulted from weathering involving liquid water during the early Hesperian (Irwin et al., 2004) and Amazonian glaciation (e.g. Head et al., 2006a,b). Due to the sensitivity and small scale of Amazonian water-related landforms, their presence provides a clear indication of conditions consistent with local surface water or ice stability. Furthermore, small perturbations away from the climate in which the specific landforms originated can cause alteration, superposition and overprinting of landforms (e.g., Marchant and Head, 2007), and thus preserve in the landscapes the record of both unique climatic conditions, and changes from these. Hence, the identification of such water ice and liquid-related equilibrium landform types can be used as a proxy for the local temperatures, precipitation levels, atmospheric pressures and wind activity prevailing when they formed; these landforms can be utilized to investigate the climate history of Mars.

In order to interpret the climate signal represented by specific landforms it is necessary to determine the processes that formed the feature, and quantify the physical limits (e.g. in terms of temperature and moisture) in which the processes operate. Terrestrial analogs are natural laboratories to investigate these processes. The key to applying terrestrial analogues to Mars lies in selecting terrestrial environments that most closely correspond to the temperature, pressure and precipitation values experienced in a particular region of Mars during the time period that is being investigated (Baker, 2001). Regions in which climate change has known to have occurred can

also be exploited to assess the limits of landform stability in response to transitional conditions (Marchant and Head, 2007).

Here we report on geomorphological investigations of Asimov crater located in Noachis Terra at 40°S, 5°E (Fig. 1 - 3) that was chosen because of its unique mid-latitude setting and its distinctive topographic configuration that optimizes an unusually wide range of insolation conditions. Within the site we have identified a multitude of distinctive ice-and-water-related landforms of various scales, including lobate debris tongues and gullies. Through the assessment of these features as potential equilibrium landforms, and investigations of their relative age and stratigraphy, we reconstruct and interpret the major local climatic conditions for this region in the Late Amazonian and assess the climate changes that have occurred. We place this analysis in the context of morphological investigations of the climatic history of the Late Amazonian conducted elsewhere on Mars (e.g., Mustard et al., 2001; Milliken et al., 2003; Head et al., 2006a,b; Morgan et al., 2009a).

To aid our interpretation of the martian landforms, we draw upon fieldwork conducted on terrestrial analogs during the Austral summer of 2006-7 and 2008-9 in the McMurdo Dry Valleys of Antarctica, a region considered to exhibit the most Mars like of conditions (e.g. Anderson et al., 1972; Gibson et al., 1983; Mahaney et al., 2001; Wentworth et al., 2005; Baker, 2001). Marchant and Head (2007) have classified the Dry Valleys into three different morphogenic zones based on unique assemblages of equilibrium landforms. Although the Antarctic Dry Valleys represent some of the most Mars-like conditions on Earth in terms of precipitation and temperature (Baker, 2001), the atmospheric composition and pressure are quite different and signal caution in making specific direct comparisons. We apply the comparisons with due attention and utilize these terrestrial analogs in terms of the nature of processes that are observed that might be applicable to both Mars and the Dry Valleys. We first outline the available data, describe the setting, and then assess the nature and significance of landforms that lead to a reconstruction of the late Amazonian climate history of the region.

2. Data Sets and Analysis

The identification of landforms within the study region and the subsequent investigation of their morphology was conducted through the use of all available image data sets. The study area was initially chosen because of the unusual setting and the large amounts of data available for this region of Mars, including full coverage of Context Imager (CTX) images (8 m/pixel) and High Resolution Stereo Camera (HRSC) images (18 m/pixel), in addition too 16 HiRISE images (0.3 m/pixel) and 73 Mars Orbital Camera (MOC) narrow angle images (1.5–12 m/pixel). The investigation was conducted using a compilation of a Geographical Information System (GIS) database comprised of the visible image datasets co-registered and overlain on digital terrain models (DTMs). The topographic data was derived from both 128 pixel/degree gridded Mars Orbiter Laser Altimeter (MOLA) data and Mars Express HRSC stereo data (200 m/pixel) derived from HRSC orbit 1932_0000. ESRI's ArcMap (9.2) provided the GIS platform, which in addition to dataset management was also used to produce slope maps and topographic profiles from the HRSC DTMs. In the areas where detailed measurements of slopes were required, individual MOLA points were extracted to provide the most accurate account of the topography. The GIS software also enabled crater diameter surveys to be conducted and size-frequency distribution plots to be compiled in order to estimate the age of surfaces. Thermal infrared data from daytime and nighttime THEMIS infrared images (band 9, 100 m/pixel) were used to characterize the thermal inertia of the surface, which in turn provides a proxy for the physical properties of the surface (such as grain sizes).

3. Geological Setting of the Study Region

Research was focused within the 80 km diameter Asimov impact crater located within the heavily cratered plains of Noachis Terra to the west of Hellas (Fig. 1). This region of Mars has been mapped within the Southern Highlands: hilly and cratered unit (Nhc) and is interpreted to be

the oldest extensively exposed surface of the planet (Scott and Carr., 1978). The Asimov crater has no identifiable ejecta deposit and its remaining rim is both discontinuous and heavily degraded (Fig. 2), suggesting that it is Noachian in age. The floor of Asimov crater is highly modified and filled and consists of a relatively flat unit that forms a surface, ~1500 m below the highest portion of the crater rim crest (~900 m below the surrounding terrain outside of the crater, Fig. 4). Depths for a fresh crater of this scale are empirically found to be >3 km according to the depth (h) to diameter (D) relationship compiled by Garvin et al (2003):

$$h = 0.36D^{0.49} \quad (1)$$

This suggests that Asimov crater has been significantly in-filled since its formation in the Noachian. The crater interior is unusual in that there is an annulus of disconnected valleys situated adjacent to the interior flanks of the crater wall. These valleys systems extend for a collective length of over 200 km, and their widths range from 3 km in the eastern edge of the basin interior to over 18 km at their widest in the south (Fig. 2 and 4). The valleys range from 200–2000 m below the current crater floor (1700–3500 m below the crater rim crest), and thus the floors of the valleys might represent a portion of the original crater floor, prior to it being filled with material. An irregular ~20 km wide depression is also present north of the center of Asimov crater (Fig. 2 and 4). The depression is connected to the surrounding valley complexes by a network of radial ~1 km wide, shallow (~20 m deep) troughs.

In the highest-resolution image datasets (MOC and HiRISE) a discrete stratigraphic layer is observed along the upper portions of the valley slopes (Figs 5-7). This unit is also apparent in the THEMIS nighttime infrared data where it correlates with a layer of higher nighttime temperatures relative to the slopes directly below (Fig. 8). This indicates that it corresponds to a material with a higher thermal inertia value, and likely represents a rock outcrop similar to others identified on Mars with THEMIS (Christensen et al., 2003). The layer appears to be actively

eroding, providing a source of scree and large boulders (> 5 meters) that are present across all slopes directly below outcrops (Figs. 5, 7). Boulder tracks, ~ 1 meter in width are visible in the highest resolution images, further suggesting that the erosion has been recent. In some sections the rock layer forms an overhang, indicating that it is more resistant than lower portions of the slopes (Fig. 6).

In areas that have HiRISE coverage it is possible to observe and study exposed outcrops of the rock layer in detail (Figs. 5-7). Prominent and pervasive polygonal fracture patterns ~10 m wide are visible along the surface of the outcrops, suggesting that erosion has exploited the surface expression of vertical rock structure (Fig. 5a). In regions where a vertical perspective of the outcrops is possible, columns of rock tens of meters tall with polygonal cross sections form the rock face (Fig. 5b). The boulders on the slopes directly below the rock layers have similar diameters to the polygonal fractures and most likely represent columns of rock that have toppled from the cliff face (Fig. 5). The structure of the rock layer closely resembles the columnar jointing present in terrestrial lava flows (Spry, 1962; Long and Wood, 1986). This type of columnar jointing is caused by contractional stresses during the cooling of the upper thermal boundary layer of a cooling lava flow (Spry, 1962; Long and Wood, 1986; Degraff et al., 1989; Grossenbacher and McDuffie, 1995). Hence, the rock layer most likely represents a lava unit that was emplaced in the interior of Asimov crater and possibly also forming the surrounding plains.

Information on the nature of the lava flow units and their cooling history can be obtained from column scale and height. Cooling from the top and bottom of the flow creates thermal boundary layers that upon solidification form columnar structures known as colonnades (Spry, 1962). Large-diameter columns are favored by slower cooling and large flows, with a slow cooling rate permitting viscous dissipation to act over a larger region (Grossenbacher and McDuffie, 1995). Toramaru and Matsumoto (2004) found that basaltic column area is inversely proportional to cooling rate. The height or thickness of columns is also related to flow thickness. The upper colonnade (well-formed basaltic columns) typically occurs in the upper third of a flow,

the lower colonnade in the lower third, with the entablature (irregular to hackly columns) forming the middle third (Long and Wood, 1986). On the basis of the width of the columns (~10 m) and the geometry of the colonnade (columns > 30 m tall) we interpret the colonnades to represent the top of a lava flow that was at least 100 m in thickness (and may have been much thicker) and that underwent very slow cooling to produce the very wide columns. Some workers interpret entablature in cooling terrestrial flows to form when the flow top is flooded by rainwater, inducing rapid cooling, convection and more irregular structure, with colonnades being favored by slower cooling by conductive heat transfer in the absence of water (Degraff et al., 1989). The tall colonnade observed on Mars in this example may be favored by the hyperarid environment and the lack of rainwater. The apparent slow cooling and large flow thickness of this unit is consistent with the mode of emplacement of the Late Noachian and Hesperian ridged plains in the region, thought to be emplaced in a flood lava mode (Peterson, 1977), fed by very large dikes (Head et al., 2006c).

In addition to the downslope movement of material from rock fall, large-scale slumping also appears to have occurred (Fig. 6). Kilometer-long arcuate fractures are present on the surface of the lava unit scarp where it is exposed at the top of the valley flanks, suggesting the development of rotational slump fractures (Fig. 6). Isolated exposures of the rock unit are found downslope of the main lava unit and are interpreted to represent slump blocks that detached from the slopes above (Fig. 6). As a result of the mass wasting operating along the valley walls, a layer of colluvium has developed below the outcropping rock layer. Therefore, it is not possible to determine unequivocally the full thickness of the lava unit. On the basis of the height of the scarp outcrop seen in Figs. 5 and 6, and the colonnade thickness estimate, we interpret the thickness of the upper flow unit to be of the order of >100 meters, about 5% of the total estimated fill in the crater interior (~ 2 km, Fig. 4). Crater fill in this region may be related to Noachian-aged age fine-grained deposits described elsewhere on Mars by Cabrol and Grin, (1999).

A fine-grained deposit is draped along the majority of surfaces within the study area and is interpreted to be related to the latitude-dependant mantle that has been observed poleward of 30° latitude on Mars (Kreslavsky and Head, 2000; Mustard et al, 2001). This mantling unit has been attributed to an atmospherically derived layer of dust-rich ice which formed during the most recent Mars ‘ice age’ when obliquity values were higher than today (Head et al., 2003). Although smooth in appearance relative to all other surfaces within Asimov crater, the mantle unit exhibits several different textures including ~10 m wide isolated pits and extended regions of cusped depressions (Fig. 7). Similar dissected morphology has also been observed in the global mantle unit in the latitude range of 30°–60° (Mustard et al., 2001; Milliken et al., 2003). This has been attributed to sublimation of water ice within the mantle caused by the current instability of near-surface ice-rich material at these latitudes (Mustard et al, 2001; Mellon and Jakosky, 1995). The abrupt change in slope associated with the surface of the steep-cliffed lava units at the top of the valleys has made it possible to observe the mantle in cross-section as it approaches the edge of the cliff (Fig. 7). This also permits estimates of the mantle thickness and internal structure to be made. Measurements were made throughout the study area where the mantle could be clearly observed, which provided an average thickness of ~20m. In some places near the margins of the deposit along the cliff faces, the appearance of distinct bands of boulders from the lava flow-unit substrate (Fig. 7) suggests that the mantle is not completely covering the lava layers, and may have been partially removed, possibly as a result of the mass wasting described above.

The formation of the deep concentric valleys clearly postdates the formation of Asimov crater and its lava plains capping fill (dating from the Noachian and Hesperian), but predates the emplacement of the Late Amazonian latitude-dependent mantle. The large-scale morphology and structure of Asimov crater provide a geologic context for the environment in which the Amazonian landforms have developed. The formation of the valleys within Asimov has been previously studied by several authors and has been attributed to the concentrated excavation of fill material along the interior walls of the crater (Schultz and Glicken, 1979; Malin and Edgett,

2000b). Malin and Edgett (2000b) acknowledge the absence of any obvious removal mechanisms to account for the loss of the crater fill material, but suggest that it may have been possible under a different climatic regime compared to current conditions. Alternatively Schultz and Glicken (1979) draw attention to similar valley depressions within younger craters elsewhere in Noachis Terra and suggest that a regional episode of endogenic activity (igneous intrusions) could have induced the collapse and removal of portions of the crater fill through the thawing of ground ice. Our observations indicate that a thick layer of lava (>100 m) was emplaced over a lower layer of less-resistant crater fill material (~ 2 km). The extent to which the top-most rock layers overhang the lower slopes argues against the lower portions of the crater fill being lava-flow-like in nature and instead suggests a weaker lithology, possibly sedimentary in origin. The estimated ~ 2 km thickness of the fill unit is consistent with sedimentary units identified within Noachian craters elsewhere on Mars (Cabrol and Grin, 1999). The occurrence of radial fractures may be a response to the load imposed by the emplacement of thick lava units over relatively weak sedimentary units. If volatiles were present within the sedimentary unit, as has been suggested for similar deposits in other martian craters (e.g. Schultz and Lutz, 1988; Fassett and Head, 2007), the release of these volatiles could have accounted for the loss of material required to form the valleys systems and central depression as suggested by Schultz and Glicken (1979). Slumping, and rock falls along the edge of the valley flanks would over time cause the widening of the valleys, and appears to be continuing into the present. These observations provide a geological and topographic context for the environments in which the late Amazonian climate-related landforms have formed.

4. Geomorphic Evidence for Liquid Water and Ice

In the following section, we describe the assemblage of ice-related landforms that we have identified within the valleys and the central depression of the crater (Fig. 3). These features are compared to landforms previously identified elsewhere on Mars in order to assess the climatic

and process conditions under which they formed. We have endeavored to determine both relative and absolute ages for the landforms where possible so that the temporal relationships between the different landform types can be established.

4.1 Gullies

The discovery of gullies on Mars has significant implications for the Amazonian climate because of the apparent role of liquid water in their formation and because of their very youthful nature (Malin and Edgett, 2000a). Gullies are typically ~1 - 2 km long and are distributed poleward of ~30° in both hemispheres, along steep slopes (>10°) within crater interiors, valleys walls and the flanks of mesas (Malin and Edgett, 2000a; Dickson et al., 2007a). The latitude dependence of gully distribution strongly suggests a climatic cause and the involvement of a volatile component in their formation. Liquid water provides the most plausible erosional agent and is consistent with the occurrence of fine-scaled fluvial features (McEwen et al., 2007), dendritic tributaries (Malin and Edgett, 2000a), levees (Malin and Edgett, 2000a) and the striking similarity between the martian gullies and terrestrial analogs formed by runoff in polar regions (e.g. Lee et al., 2001; Costard et al., 2002; Morgan et al., 2008).

Models of gully formation require an explanation of how pressure/temperature conditions optimal for melting and flow can be achieved on the surface of Mars (albeit briefly) during the Late Amazonian in order to permit the occurrence of liquid water. Initial models invoked the rapid release of groundwater discharge from perched aquifers (Malin and Edgett, 2000a; Heldmann and Mellon, 2004). Further considerations regarding the current/recent metastability of liquid water on the surface of Mars generated alternative explanations, including atmospherically deposited sources of water (Hecht, 2002; Costard et al., 2002; Christensen, 2003; Dickson et al., 2007a). Many of the models propose that gullies formed during periods of higher obliquity when H₂O ice is redistributed from the poles to the mid-latitudes (Costard et al., 2002;

Milliken et al., 2003; Christensen, 2003; Dickson et al., 2007a; Head et al., 2003; Morgan et al., 2009b). The release of water and CO₂ ice from the poles via sublimation under these conditions would have also increased the surface atmospheric pressure (Kieffer and Zent, 1992) and water vapor content of the atmosphere (Mellon and Jakosky, 1995), and thus further encourage gully development. Therefore, the study of gullies can be used to identify microclimate zones where ice rich deposits were able to accumulate and melt during recent martian history, and thus provide a proxy for the past climates and climate change.

4.1.1 Location and Morphology

Gullies are present along every slope within the valley systems and the central depression; their morphology, however, is highly dependent on slope aspect (Fig 9). Differences in gully morphology in this area has been assessed by Morgan et al. (2009b), who concluded that the gullies form as a result of top-down heating and melting of snow; their model is consistent with other gully studies that suggest an atmospherically deposited water source (Hecht, 2002; Costard et al., 2002; Christensen, 2003; Dickson et al., 2007a). Here, and in later sections, we expand upon the Morgan et al (2009b) work and assess gully relationships with other ice-related landforms.

The poleward-facing gullies within the study area (Fig 9a) are highly developed and heavily incised into the slopes of the valley walls. The gullies typically consist of multiple tributaries which originate in alcoves within the outcropping cap rock unit at the top of the slope and merge down slope to form single ~100 m wide channels that open up into fan deposits. At MOC and HiRISE resolution (sub-meter) it is possible to identify individual 5 m wide sinuous channels within the main channel (Fig. 9c); these exhibit streamlined islands and potential terraces, similar to those identified in gullies elsewhere on Mars (McEwen et al., 2007). The gullies present on north (equatorward-facing) slopes (Fig. 9b) display morphologic and scale

differences relative to the poleward-facing gullies; these originate in ~500 m wide amphitheater-shaped alcoves that are eroded into cap the rock unit. Multiple, relatively straight channels originate from the apex at the bottom of the alcoves and extend for 500 - 1000 m until they open up into fan deposits (Fig. 9b). The equator-facing gully channels lack the dendritic system of tributaries present in the poleward facing gullies and are significantly less incised into the valley wall. The occurrence of gullies on isolated slopes along the narrow walls (300 m wide) that divide the heads of the disconnected valleys (Fig. 10), argues against the occurrence of perched aquifers providing the gully water source. Morgan et al. (2009b) underlined this observation as an argument in support of an external source of water, and suggested that the melt of atmospherically deposited snow was the most likely candidate for the formation of these gullies.

Smaller gully forms only resolvable at HiRISE resolutions are also present along slopes within the lower portions of the valleys. There are two distinct morphological types. The first type is located along the side of boulder-covered mounds within the valley floors (Fig. 11). These exhibit the three morphological components (alcove, channel and depositional fan) used to define gullies on Mars by Malin and Edgett (2000a), although they are much smaller than the other larger gullies within Asimov crater (channel lengths are ~200 m, Fig. 11). The alcoves are 20 m wide and are positioned adjacent to each other in single rows cut into the flanks of the mounds. Thin < 1 m wide channels emerge from the alcoves and maintain straight courses except for gentle meanders around boulders (Fig 11). The slope between the boulders appears smooth and frequently displays fine scaled aeolian ripples (wavelength of 1 m) suggesting the gullies are cut into a dust rich layer.

The second gully type is located along the side of a dune-covered mound within the southern valley complex (Fig. 12). The dunes are linear with a wavelength of ~ 3m. Smaller ripples with a wavelength of ~ 1 m are present orthogonal to the dune crests, suggesting that the dominant winds that formed the dunes were bidirectional, and perpendicular to each other. The gullies are only comprised of channels, and do not exhibit well-developed alcoves or depositional

fans. The channels are ~200 m long, 3 m wide, are bounded by levees, and display largely linear courses except for small-tight meanders (Fig. 12). The level of incision of individual adjacent channels differs, suggesting that some of the channels were abandoned in favor of other courses. The gullies are orientated in the same direction as the dunes and are of a similar width, suggesting that the dune structure strongly influenced the gully formation.

The crisp and unweathered nature of these two groups of gullies, and their small scale, supports the interpretation that they are very young. It is therefore possible that they formed at the same time as the other larger scale gullies, or alternatively they may result from more recent activity. If they were formed by snowmelt as has been proposed for the larger gullies by Morgan et al (2009b), then smaller volumes of melt would have been required. One explanation for the origin of these smaller gullies may relate to the localized climate effects associated with elevation. The apices of the larger gully alcoves (i.e. the uppermost section of the channels) are at an altitude of around 500 m, whereas the smaller gullies have been found at ~1000 m. Both of these values are within the elevation range that gullies occur on Mars (-5000 to + 3000 m, Dickson et al., 2007a). Nevertheless the 1500 m elevation range between these two gully groups within the study area may be important; the atmospheric lapse rate on Mars is low, and so there is no appreciable surface temperature gradient between these two altitudes. However, the atmospheric pressure decrease with elevation is significant with regard to water stability. Lobitz et al (2001) derived an expression for the relationship between elevation and surface pressure on Mars, constrained by the Viking 2 lander pressure measurements:

$$P(z, L_s) = P_{VL2}(L_s) \exp(-(z - z_{VL2})/H) \quad (2)$$

where H is the scale height for Mars, 10.8 km, $P_{VL2}(L_s)$ is the surface pressure at the Viking 2 Lander site at solar longitude L_s (fit to a polynomial curve), and z_{VL2} is the MOLA-derived altitude at the Viking 2 Lander site (-4 km). We used this in conjunction with the LMD

1D GCM model of Mars surface temperatures to plot the annual difference in climate (average pressure, maximum temperatures) between the two elevations of gully alcoves in temperature and pressure space (Fig. 13). During limited periods of the day in the spring, the small-scale gully lower elevation experiences conditions above the triple point. Further modeling is required to determine if these conditions would be enough to permit melting to form small gullies.

4.1.2 Comparisons With other Martian Gullies

Aspect-dependent gully morphology has been documented elsewhere on Mars; Hale crater, which also displays the same relationship between well-developed, deeply-incised, pole-facing gullies and smaller, more subdued gully-like forms on equator facing slopes (Riess et al., 2009). The origin of the equator-facing features in Hale is unclear (they do not have incised channels, and are superimposed by impact craters). This suggests that they might have formed shortly after the formation of Hale and thus might be related to the impact event itself in a manner similar to the proposed formation of Mojave crater fluvial features (Williams et al., 2004). The gullies in the study area are also significantly younger than the Noachian aged Asimov crater and the crater fill material into which they are eroded, and so are not related to the original impact event. Despite the differences in gully morphology within the study site, all the large-scale gullies exhibit the basic morphologic characteristic of the other mid-latitude gully deposits originally documented by Malin and Edgett (2000a), supporting the argument of Morgan et al. (2009b) that they were formed during higher obliquity excursions in a manner similar to that proposed for gullies elsewhere on Mars (Costard et al., 2002; Christensen, 2003; Dickson et al., 2007a; Head et al., 2008).

The smaller scale gullies located close to the valley floor (Fig 11 and 12) are only observable at HiRISE resolution and thus similar examples have not been documented in gully studies based only on MOC data. The prominence of channels within these gully systems implies

that water was involved in their formation and they may simply be scaled-down versions of ‘typical’ martian gullies documented by Malin and Edgett (2000a). The gullies on the sand dunes however, are morphologically similar to larger-scale gully features found on dunes in other regions of Mars. Dune gullies in the study area are compared to those in Russell Crater (55°S, 13°E) in figure. 12. Both of these gully types have relatively long and thin linear channels and no obvious fan deposits, though the Russell crater gullies are ~2.5 km long and originate in poorly developed alcoves (Mangold et al., 2003). The Russell dune gullies have been attributed to debris flows initiated by liquid water (Mangold et al., 2003; Reiss and Jaumann, 2004; Miyamoto et al., 2004; Vedic et al., 2008), and a similar process may have formed the study area gullies. Their similar morphology, despite the large difference in scale, may be the result of the two gully sets forming in finer grain-sized, largely unconsolidated deposits compared to the larger gullies located along the slopes of interior crater walls and valleys. The width of the dune gullies in the study area appears to be controlled by the wavelength of the dunes on which they have formed (Fig. 12), potentially demonstrating the greater influence of dune surface structure at smaller scales.

4.1.3 Age of the Gullies

Constraining the age of gullies is difficult because they cover small areas and are limited to steep slopes, which are prone to failure and thus are poor recorders of impact craters. Nevertheless, through the dating of a dune field that has been superposed by gully fans (Reiss et al., 2004), and through the identification and dating of a primary impact crater that has generated secondary impacts on a gully fan (Schon et al., 2009), the most recent episode of gully activity is likely to have occurred ~500 Ka to 3 Ma for two locations on Mars. Despite the inability to utilize crater counts to provide absolute ages for individual gullies within the study area, crater counts can still be used to compare the relative ages between the different morphological gully types. A

survey conducted of all available MOC (22) and HiRISE (8) images revealed that significantly more craters are superimposed on the fans and channels of equator facing slopes than pole facing slopes, including 6 craters with a diameter > 100 m. This suggests that the pole-facing gullies have been the most recently active, which is consistent with them being more incised and extensively developed (in terms of tributaries and fine scaled fluvial features) than the simpler equator facing slopes.

Morgan et al. (2009b), in their investigation of gully formation within the study site, used a 1D version of the LMD/GCM (Forget et al., 1999) to show that the range in gully morphology is correlated with aspect-dependent surface temperatures occurring during higher obliquities. Their model predicts that the pole-facing gullies could have formed during high obliquity (35°) excursions that occurred periodically over the last 2.5 Ma (Laskar et al., 2004), with the most recent activity potentially occurring >500 Ka. This is consistent with the age estimates of gully activity by Reiss et al. (2004) and Schon et al. (2009) (see table. 1). However, for equator-facing gullies, Morgan et al (2009b) found that they could only be formed during the highest obliquity environments that occurred prior to 4 Ma ($\sim 45^\circ$); thus, they interpret the cause of the different morphologies to be due to: (1) the longer period of inactivity (potentially 5 Ma for the equator facing gullies, compared to ~ 500 Ka for the polar facing gullies) and (2) the limited period of activity during each obliquity oscillation prior to 5 Ma (obliquities of 45° were maintained for 20,000 yrs during each 110 ka obliquity cycle) relative to that of the polar facing gullies (obliquities > 35 were maintained for 70,000 yrs during each 110 ka obliquity cycle).

4.2 Viscous Lobe Features

4.2.1 Location and Morphology

Throughout the valley systems are a series of tongue-shaped, debris-rich lobes (Fig. 14) that display surface morphological features indicative of viscous flow. We term these lobate

debris tongues (LDT), describe them, and compare them to other similar features on Mars and Earth. These landforms originate at the heads of valleys (Fig. 14), extend for several kilometers, and terminate on the valley floor. They exhibit convex-upward topographic profiles and surface fold-like compressional ridges (Fig. 14c and inset), suggesting that they have experienced ductile deformation during downslope flow. In some cases smaller scale tributary flows have contributed to the formation of large-scale lobate features (Fig. 14b) indicating that integrated flow from several sources has occurred. All of these factors suggest that the lobate debris tongues may contain ice that may have acted to facilitate the downslope creep of debris. Therefore, investigating these features became a major focus of our research.

Due to the relatively small size of the lobate debris tongue (3 – 5 km in length) it is difficult to make accurate measurements of their topography as the along-track spacing of MOLA shots results in a maximum coverage of 5-10 points in a north-south direction (along-track direction). This number can be significantly less for east-west measurements because of widely spaced orbital tracks. Nevertheless, by extracting points from multiple adjacent orbital tracks, we were able to construct a topographic profile for the specific LDT that occupies the eastern head of the southern-most valley (Fig. 15). Extrapolating the topography of the valley floor below the LDT indicates a maximum thickness for the deposit of ~250 m.

The lobate debris tongue surfaces are covered with boulders up to ~10 meters in diameter. These appear similar in albedo and shape to those originating from the rock layer at the top of the slope (Figs. 5, 7) and this suggests that the lobes are largely composed of debris shed from these layers. Many of the large boulders (>5m) within the valleys are surrounded by circular depressions (moats) that are up to ~2 m in width (Fig. 16a,b,c). Width of the moats appears to be a function of the size of the block (Fig. 16). Erosional scouring associated with eolian activity appears unlikely to have caused the moats features, as one would expect the moats to be orientated along the dominant wind direction and to exhibit depositional tails on the lee side of the boulders. The large blocks appear to be comprised of solid rock derived from the flow layers

at the wall summit, and are dark relative to the surrounding finer-grained sediments and dust. Thus, as an alternative explanation, we interpret the moats to have been formed by local preferential sublimation of near-surface ice due to the higher thermal inertia and lower albedo of the boulders. In this model, the boulders will act as thermal capacitors, absorbing shortwave solar energy and emitting long-wave energy, increasing the temperature of surrounding ground surface, and enhancing the effects of sublimation. To our knowledge, such features have not been reported in the literature, and if further research supports this interpretation, such occurrences may provide a useful proxy for the occurrence of near-surface ground ice.

Larger-scale topographic features are also apparent along the surface of the LDT deposits. Parallel elongated ~200 m wide troughs are present along the central axis of some of the lobate debris tongues (Fig. 16b). The troughs may be a product of deflation processes and might represent the lowering of the surface due to the loss of ice through sublimation. Smaller scale, linear indentations tens of meters in width are located along the upper, steeper portions of the LDT (Fig. 14). The indentations may have formed as the result of strain induced by brittle deformation within the most elevated portions of the lobes.

These texture and features associated with the surface of the LDT are also found along the floors of the narrower (< 5 km) sections of the valleys. To the north of the valley shown in Figure 14a (see Fig. 16a for context), larger troughs are present that form a series of 150 m wide concentric terraces (Fig. 16d). Boulders surrounded by moats are also present in association with the troughs. The presence of features interpreted to represent the occurrence of near-surface ice suggests that portions of the valley floors may contain ice-rich material.

Gullies are found along all slopes from which the lobate debris tongues originate and in some cases the gully fan deposits are emplaced on top of the surface of the upper portions of the LDT (Figs. 14a,c, 15). At the site of the LDT in Fig. 14a the termini of the gully channels and their fans curve downslope following the direction of advance of the lobe. Despite this general trend, there is one area in the southern valley (Fig. 3) where a LDT appears to be superimposed

on top of a gully (Fig. 17, 18). Along the lower portions of the southern valley, a somewhat degraded gully channel ~40 m wide emanates from below the lobe front of the LDT (Fig. 18). There is a sharp transition between the LDT termini and the upper portions of the gully channel, which suggests that the two features are distinct and that the LDT is not the source of the channel. Understanding the processes by which ice was emplaced to form the LDT, the volume and fate of the ice involved, and the nature of the relationship between the LDT and the gullies is thus essential to the reconstruction of the local climate history of the region.

4.2.2 Comparisons With other Martian Lobate Flow Features

Viscous flow features of various scales that are interpreted to be comprised of ice-rich materials have been identified elsewhere on the martian surface (Squyres, 1978). Two types of viscous flow features that also exhibit lobe-like morphology: lobate debris aprons (LDA) and lineated valley fill (LVF). These are found at ~30°-50° latitude in both hemispheres, and have been interpreted to be either ice-assisted creep of talus (e.g., Squyres, 1978), or deposits from debris-covered glaciers (e.g. Head et al, 2006a,b). Lobate debris aprons form extensive deposits around isolated mesas (Squyres, 1979; Pierce and Crown, 2003) and appear to be genetically related to lineated valley fill, which forms large-scale integrated patterns and extends for hundreds of kilometers through fretted valleys (Head et al., 2006a,b; Morgan et al., 2009a). However, unlike LDA and LVF, the LDT are an order of magnitude smaller and display different surface textures, consisting of large boulders and linear troughs. Although LVF/LDA deposits show evidence of flow structures, they also have pits and buttes that are interpreted to be related to large-scale sublimation processes (Mangold, 2003), features that are not observed on the LDT. Additional LDT traits not typically shared by LVF/LDA deposits include 1) the occurrence of large numbers of gullies in the vicinity of the tongue, and 2) a paucity of superposed impact craters. LDA/LVF have numerous superposed craters and ages in excess of 100 million years

(e.g., Mangold, 2003; Kress and Head, 2009).

Milliken et al. (2003) have reported on the presence of younger viscous flow features that are much more similar in scale to the LDT than are the LDA and LVF, and these show a close association with gullies (Milliken et al., 2003; Head et al., 2003; Arfstom and Hartmann, 2005; Reiss et al., 2009; Head et al., 2008). Milliken et al (2003) identified ~10 m thick viscous flow features (VFF) throughout the 30 – 60° latitude bands which they interpreted to be related to be remnants of ice-rich dust deposits associated with the climate-related, latitude-dependent mantle (Mustard et al, 2001; Head et al., 2003). In a manner very similar to the relationship between LDT and gullies in the study site, Milliken et al. (2003) documented gullies that cut through the VFF, suggesting a younger age for the gullies and that the gullies may have been formed by the melting of ice associated with the latitudinal dependent mantle. These VFF are significantly thinner than the LDT and contain fewer boulders, most likely due to the VFF being composed of more dust-rich material compared to the rocky material that constitutes the LDT. In many locations gullies are also found in close proximity to arcuate lobate features that form around the gully fans (Milliken et al., 2003; Head et al., 2008). These features are interpreted to be ‘beheaded’ glacial systems, which have lost ice from within former accumulation zones (Head et al., 2008). These features differ in shape, setting and morphology relative to the LDT. LDT lack the thick distal deposits, the beheaded structures, and occur along valley walls, not in crater interiors (e.g., Head et al, 2008).

Determining the amount of ice present in debris-rich viscous flow features from morphology alone is a difficult challenge on both Earth and Mars (Whalley and Azizi, 2003; Mangold, 2003; Li et al., 2005; Pierce and Crown, 2003). Several issues further complicate this determination: 1) How much ice is required to cause mobilization of rock debris to produce flow-like features? 2) In what configuration did this ice occur during formation of the features (e.g., ice in pore-space of debris fans or debris covered glaciers)? 3) Do the currently observed features still contain the ice that existed at the time of their formation, or have they lost some or all of this

ice? 4) What nomenclature best applies to the description of these features (e.g., Whalley and Azizi, 2003)? Recent work in the Antarctic Dry Valleys has helped to clarify some of these questions. Documentation of the characteristics of debris-covered glaciers in Beacon Valley, Antarctica (Marchant et al., 2002), and criteria for distinguishing these from other viscous flow features formed there (Marchant and Head, 2007), leads to a further understanding of the range of candidates and an appropriate nomenclature. We prefer to use the term "debris-covered glacier" where geological evidence supports the presence of subsurface ice that is significantly in excess of debris pore space. Such evidence includes flow textures and structures (Head et al., 2006a,b), large sublimation pits and unusual superposed craters (Kress and Head, 2008), surface polygon structures and their geometry (Levy et al., 2006), topographic profiles (Li et al., 2005), flow distances, detection of abundant subsurface ice by radar (Holt et al., 2008; Plaut et al., 2009) or seismic surveys (Shean et al., 2007b), and experimental and rheological considerations (Mangold and Allemand, 2001; Mangold, 2003). We prefer to use the term "rock glacier" where rocky lobate viscous flow features do not meet these criteria. By these definitions, "rock glaciers" could be either talus deposits mobilized by interstitial ice, or "debris-covered glaciers" that have lost their ice due to sublimation and melting, and deflated to their currently observed configuration. For example, the occurrence of linear troughs in the lobate debris tongues (Fig. 16) suggests that the LDT have undergone deflation from loss of ice due to sublimation and thus they may not be currently active.

Understanding the source of ice associated with the formation of the LDT and establishing potential sources of water that produced the gully systems is fundamental to deriving the climate conditions under which they both formed. Groundwater injection into debris fans has been proposed as a potential bottom-up source of ice for lobate flow features on Mars (e.g. Mangold and Allemand, 2001.). For example, if water was derived from the subsurface in sufficient volumes, it could saturate the pore spaces of debris and/or produce ice lenses; this might enable ice-assisted creep in a manner similar to terrestrial rock glaciers (Whalley and

Azizi, 2003). For such a mechanism to operate in the study area on Mars, perched aquifers in communication with the walls of the valleys would be required. The occurrence of gullies and LDT along isolated ridges (Fig. 10), however, is inconsistent with groundwater and instead argues for an external, atmospherically linked, top-down, source of ice (see section 4.1.1). A debris-covered glacial origin for the lobate debris tongue would require conditions favorable for significant precipitation of snow and ice, and its accumulation and flow, and pressures and temperature conditions consistent with the stability of surface ice.

4.2.3 Age of Viscous Flow Features

The superposition of gullies along the upper portions of the LDT indicates that the gullies are predominantly younger than the LDT. Thus a period of gully formation is interpreted to have followed the emplacement of the LDT. The larger scale and lower slope of the surface of the LDT relative to the gully systems makes it possible to make crater count surveys to provide age estimates. Crater counts and crater size-frequency distributions were compiled for a lobate debris tongue in order to derive an age for the surface of the feature. We also noted the position of each crater in order to investigate areas with anomalously lower numbers of craters that might signify relatively recent activity or modification. Figure 19 highlights the spatial distribution of impact craters on the surface of the LDT in the southern portion of the north-eastern valley complex (Figs. 3, 16a). One hundred craters were identified on the surface of the LDT; the largest crater had a diameter of ~30 m. The craters are not evenly distributed, and do not appear to be random, as would be expected if the whole surface was of the same age. They also do not appear to be tightly clustered, as is often the case when single projectiles brake up during their travel through the atmosphere and produce a cluster of craters (Malin et al., 2006). Specifically, there appears to be a distinct absence of craters along the snout and central portions of the LDT and along the linear troughs (Fig. 19). This may result from recent modification as a result of deflation.

The size-frequency distribution of the craters were plotted according to Hartmann (2005) isochrons (Fig. 20). If we assume the rollover in smaller craters (< 10 m D) is due to a modification of the surface (possibly from the same processes that caused the uneven spatial distribution of the craters), the plot fits close to the 10 Ma Hartmann isochrons. However, because the LDT lies between the two steep (16°), 500 m high slopes we need to consider the shielding effects caused by the limited sky exposed to the surface of the LDT, which would act to filter out oblique impacts. The geometry of the valley cross section would prevent impacts greater than 35° impacting a point on the surface of the LDT at an equal distance from each valley wall (Fig. 21). The probability of a projectile impacting the surface between θ and $\theta + d\theta$ is directly proportional to:

$$dP = \sin \theta \cos \theta d\theta \quad (3)$$

This dictates that 32% of the impacts will be blocked. However, if we also take into consideration that impact angles less than 15° are likely to generate non-circular craters (and so would not have been counted in the crater survey) (Pierazzo and Melosh, 2000), then the LDT is likely to have experienced 35% less impacts than if it was on a fully exposed surface. To simulate the effects of the sky filtering we added 35% craters to each crater diameter bin and replotted the results. This increases the estimated age of LDT to above the 10 Ma isochron (Fig. 20). Also if we consider the uneven distribution of craters on the LDT surface (Fig. 19), the age above relates to the modification of the surface, so the actual age of the formation of the LDT is likely to be older than this.

4.3 Structural Ridges

4.3.1 Location and Morphology

Elongated, linear ridges are present along the lower portions of the southern slopes in the southern most valley system (Figs. 3, 22). The ridges are situated close to, but below the main transition in slope between the upper, steeper slopes (20° to $>30^{\circ}$) and the lower more gently sloping ($<10^{\circ}$) portions of the valley walls (Fig. 2). The ridges extend discontinuously along the southern wall for ~ 45 km. The small scale of the ridges (about 100 m wide) prevents the precise measurement of their heights using topographic MOLA profiles; shadow measurements show them to be of the order of tens of meters high. Analysis of the surfaces of the ridges using high-resolution images shows them to consist either of a series of blocky hummocks (~ 10 m across, Fig. 23a) or elongated steep crests (Fig. 23b,c) that sit on top of a broad elevated structure ~ 100 m wide. Both ridge types are interspersed with boulders, which in some locations appear to be undergoing exhumation directly out of the surface of the ridge. In the eastern end of the southern valley (Fig. 3), the ridges are present along the termini of a small LDT-like lobe (Fig. 18). The LDT has a characteristic distinct lobe-shaped front, but appears more deflated than the examples shown in Fig. 14.

These features are morphologically similar to terrestrial moraines (For further details on moraines: Bennett and Glasser, 1996; Benn and Evans, 1998). Moraines on Earth represent a range of depositional features formed by the concentration of debris that has been transported superglacially, subglacially or englacially to the margins of a glacier system (Benn and Evans, 1998). The association of the ridges with the distal margin of a degraded LDT further supports their interpretation as moraines and suggest that they were formed by the transport of debris downslope by the LDT. Therefore, these ridges were investigated further because of their potential climatic significance and because of the additional information these features can provide regarding the origin and degradation of LDT deposits.

4.3.2 Comparisons With other Martian Structural Ridge Features

Potential moraines of a similar scale have been identified in association with small-scale LVF lobes in the Protonilus Mensae-Coloe Fossae region, and have been interpreted to represent recent glacial highstands formed during Late Amazonian periods of renewed glaciation (Dickson et al., 2008). Similar features have also been discovered in the southern hemisphere, around the bases of massifs in Bond crater, and are interpreted to have formed by glacial activity in the region (Reiss et al., 2009). Arfstrom and Hartmann (2005) documented moraine-like ridges along the foot of gullied intercrater walls across the southern mid-latitudes. However, protalus ramparts that are formed on Earth by the downslope movement of mass-wasting products along perennial snowbanks can also develop into ridges along the base of slopes. Arfstrom and Hartmann (2005) argue that the ~1 km distance of the moraine-like ridges from debris sources and the 20 - 30° gradient of the slopes below which they form is not compatible with protalus rampart development, which on Earth is limited to 50-100 m from the base of near-vertical slopes (60-90°) (Ballantyne and Harris, 1994). They therefore, concluded that the moraine-like ridges were of glacial origin.

This same argument also holds true for the ridges in the study area. The occurrence of moraine-like ridges in association with lobate flow features, similar to what has been observed in the eastern section of the southern valley (Fig. 6), has been reported within crater wall environments (Arfstrom and Hartmann, 2005) and small scale LVF lobes along the dichotomy boundary (Dickson et al., 2008). In both cases the ridges were interpreted to be terminal moraines. We interpret the ridge structures on the LDT as resulting from local high concentrations of material that has been transported from up slope along the surface or within the body of the LDT (possibly along flow lines). The ridges elsewhere in the southern valley (Fig. 22, 23) might also mark the extent of other LDT, which have since completely disappeared.

4.3.3 Ages of Structural Ridges

Craters up to 15 m diameter in size are present on the surface of the ridges, but it is difficult to constrain an age from the crater size-frequency distribution because of insufficient surface area on which to conduct crater counts. The inferred origin of the ridges as moraines formed by the degradation of LDT deposits implies that their age represents climatic change associated with the demise of LDT and thus the age of the ridges is likely to be close to the >10 Ma old estimate of the surface modification age of the LDT (Fig. 20; section 4.2.3).

4.4 Summary of Observations

Three distinct types of Amazonian landforms were identified within the study region that demonstrate morphological traits that are suggestive of being formed by the action of surficial deposits of snow/ice. These features are; LDT, gullies and linear ridges. Surface features indicative of viscous flow (fold-like compressional ridges, convex-upward topographical profiles) along the bodies of the LDT are similar to other features on Mars interpreted to be ice rich, and suggests that significant volumes of ice were deposited within the valleys prior to 10 Ma (current surface age estimate of LDT, Fig. 20). The morphological similarities between the linear ridges and terrestrial moraines and their spatial association with the distal margins of the LDT argues for a climatic shift resulting in the demise of LDT features through the loss of ice and the associated deposition of debris along the southern valley complex (Fig. 22). Gullies have eroded into the surface of the upper portions of LDT deposits and are interpreted to have originated during a climatic period that postdated the initial formation and subsequent degradation of the LDT.

In order to understand the climatic history responsible for the formation of the landform assemblages present in the study area requires: 1) The volumes of ice required to generate each landform type (and thus the levels of precipitation and surface temperatures required to facilitate the deposition of the ice) to be established. 2) The stability of each landform in the presence of

climatic transitions to be understood (thus we must be able to distinguish between relict landforms or components of the landforms from features that are in equilibrium with the current climate). Terrestrial analogs offer a powerful tool to address these questions as the processes responsible for the formation of similar landforms can be observed in real time and the effects of overprinting of different landform assemblages (such as the gullies on top of the LDT Fig. 14, 15) and the structural integrity of these features during active climatic changes, can be assessed.

5. Terrestrial Analogs in the McMurdo Dry Valleys of Antarctica

Field research was conducted in the McMurdo Dry Valleys of Antarctica (ADV) on landforms morphologically analogous to the features identified within the study region in order to: 1) conduct in situ investigations into the features, their current and past ice content, and where possible their three-dimensional structure, 2) place further constraints on the environmental conditions (e.g. maximum-minimum temperatures, precipitation levels) under which the landforms may have formed, 3) test the physical limits and the extent of modification these features experience in the presence of climatic change. The Antarctic Dry Valleys represent an ice-free region of the Transantarctic Mountains, situated between the East Antarctic Ice Sheet and the Ross Sea (Fig. 24). The ADV are a hyperarid, cold polar desert in which sublimation exceeds precipitation (Marchant and Head, 2007) and has thus long been held to provide one of the closest terrestrial analogs for the current martian conditions (e.g. Anderson et al., 1972; Gibson et al., 1983). Marchant and Head (2007) subdivided the ADV into three morphogenic zones based on unique assemblages of landforms that are in climatic equilibrium with their local micro-environmental conditions. These landform groups are sensitive to local precipitation levels and temperatures, both of which decrease as a function of distance inland from the Ross Sea (Fig. 24) and with the corresponding increase in elevation. Marchant and Head (2007) used the identification of currently stable landforms in these different microenvironments, and the

assessment of evidence for the superposition and migration of microclimatic landform types into adjacent areas, to analyze recent trends in climatic change throughout the ADV. In order to concentrate our efforts within the climatic conditions most representative of Late Amazonian Mars (Marchant and Head, 2007), we conducted our research in the driest and coldest portion of the ADV within the stable upland zone (SUZ) and the most elevated regions of the intermediate mixed zone (IMZ).

5.1 Upland Stable Zone (USZ)

Summertime mean air temperatures in the USZ are about -10°C , and the ground rarely experiences temperatures in excess of the freezing point, preventing the formation of an active layer (upper portion of the surface that experiences ground thaw; Marchant and Head, 2007). Sublimation is the dominant form of mass loss from any surficial deposits of snow or ice, resulting in negligible amounts of meltwater; this prohibits any significant surface runoff. Research was conducted within Beacon Valley, a 12 km long valley system with multiple adjoining tributary valleys, located within the Quartermain Mountain Range. Beacon Valley has previously been the focus of Mars research because it contains slow moving (<40 mm/yr; Rignot et al., 2002) integrated systems of debris-covered glaciers (Marchant et al., 2002). These debris-covered glaciers are morphologically distinct from the martian LDT and have been used as an analog for LVF/LDA systems (e.g. Head et al, 2006a,b). Along the northern wall of Beacon Valley, however, there are smaller-scale alpine glacial features that are similar to the LDT. The northwestern wall of Beacon (Fig. 25) is physically similar to the martian study site (Fig 2,3), and consists of steep walls ($>25^{\circ}$) that are capped by columnar-jointed Ferrar Dolerite sills, 100 m thick. Relatively small-scale glaciers (~ 0.5 - 1 km long) have formed on the valley walls. The low levels of precipitation and the strong katabatic winds that blow down from the plateau result in significant ablation across the glacial surfaces, exposing blue glacial ice. The glaciers are not

directly fed by precipitation, but by avalanches generated by the collapse of ice along the edge of the plateau icefield above (Fig. 25). In addition to ice, rock-fall debris is also supplied to the surface of the glaciers from the intervening dolerite cliffs. Ice-cored moraines 50 m wide fringe the present ice margin (Fig. 25). Beyond these moraines are ice-cored lobes 70 m in length, which presumably represent downslope flow of older (distal) ice-cored moraines (Fig. 25, 26). The surface of these lobes consist of poorly sorted angular boulders of dolerite 0.1 – 1 m in diameter that are stacked at the angle of repose ($>40^\circ$), suggesting that the lobes are currently active (Fig 26).

The small-scale glaciers and ice-cored lobes share several similarities to the LDT identified in the martian study area (Fig. 2,3), including: broad lobate morphology (compare Fig. 14 with Fig. 25b), convex-up topographic profiles (compare Fig. 15 with Fig. 25a), debris covered surfaces comprised of angular boulders (compare Fig. 16 with Fig. 26b), and are located within a physical setting consisting of steep valley walls capped with near vertical resistant cliffs with igneous rock outcrops (compare Figs. 2, 5, 14 with Fig. 25). Thus, these Beacon valley glaciers may provide some insights into the mode of formation and evolution of the LDT in the study area on Mars. The development of ice-cored lobes in the Beacon alpine glaciers may be analogous to the manner in which the LDT developed into full debris-covered features. This ADV example demonstrates that within a hyperarid, cold desert environment, ice rich viscous flow features can form provided that there: (1) is a source of ice (such as an ice field), (2) is a debris supply that can provide insulation from the atmosphere to prevent sublimation (either by the debris directly falling on the surface of the ice, or through the formation of a sublimation lag) and (3) are steep slopes to induce the down slope movement of the resulting ice-rich debris.

5.2 *Intermediate Mixed Zone (IMZ)*

Gullies are the most prevalent and stratigraphically youngest landform within the martian

study site, and their presence argues for the occurrence of liquid water during their formation. In order to investigate analogs for the formation of liquid water under Amazonian martian conditions, and the geomorphic effect of the runoff produced, research was conducted within the IMZ. We concentrated our efforts within the most elevated (and hence driest) portion of the IMZ situated within South Fork (SF), an 8 km long valley that forms the southern extent of Upper Wright Valley (Fig. 24), where it is divided by Dias plateau. South Fork is ~2 km wide and is flanked to the north and the south by steep slopes ($>30^\circ$). This region marks the most inland extent of active fluvial features within the ADV and is most analogous to Mars during Amazonian periods of high obliquity in terms of precipitation and maximum temperature values in excess of the melting point of water (Fig. 27). Unlike other gullied regions of the ADV, there are no significant adjacent glacial systems to supply meltwater, or to feed smaller-scale glaciers (Fig. 24).

Previous work within South Fork has shown that despite the low precipitation (<10 mm per year) and average mean summer air temperatures of below 0°C , gully activity can occur from the melting of both perennial snow banks (in the gully alcoves) and annually deposited snow packs within gully channels (Morgan et al., 2007, 2008). Wind action can redistribute the small amount of annual snow that falls in this environment, and cause it to be trapped in topographic lows, such as gully channels. In this manner, snow packs tens of centimeters thick can build up in gully channels. Melting of the snow can occur daily during periods of direct solar insolation under cloud-free conditions. The runoff generated by the snowmelt is capable of eroding small-scale fluvial features within the gully channels and fans: these gullies have been proposed as terrestrial analogs for gully activity on Mars (Dickson et al., 2007b; Levy et al., 2007; Morgan et al., 2007; 2008).

Within South Fork we also investigated a lobate flow-like feature that is morphologically very similar to that of the martian lobe (Fig. 28). The lobe extends from the up-valley side of South Fork and terminates along the western shore of Don Juan Pond. The snout of the lobe

exhibits a convex up profile which levels off around ~20 m above the valley floor (Fig. 29, 30). Gully systems are located along the southern wall of the valley. Several of these exhibit channel termini and fans that bend into the down-slope direction of the tongue itself in a similar fashion to those of the martian LDT (Fig 28). Somewhat larger channel systems are also present along the center of lobe; these terminate at its snout in a series of delta systems adjacent to Don Juan Pond.

The lobe is referred to in the literature as a 'rock glacier' (Hassinger and Mayewski, 1983; Bockheim and McLeod, 2008), but little specific research has been devoted to the composition and formation of the feature. Hassinger and Mayewski (1983) argue that the lack of a steep terminus and its convex-up cross sectional profile are indicative of the lobe currently being inactive (i.e. it has undergone viscous relaxation). A seismic refraction station set at the terminus found no definitive evidence to suggest that primary ice was present, but instead interpreted the interior to consist of interstitial ice within pore spaces (Hassinger and Mayewski, 1983). This was further confirmed by a 100 m seismic line set up further up the body of the tongue, which found no evidence of a strong reflection associated with primary ice layers (Shean personal communication). The age of the lobe and other 'rock glaciers' within the region have been correlated with the Taylor III drift, 208-335 ka (Brook et al, 1993) based on comparable surface weathering characteristics (Bockheim and McLeod, 2008). Although this feature meets our criteria for being termed a "rock glacier", it is not yet clear if the feature formed in this manner, or whether it is a "debris-covered glacier" subsequently deflated by sublimation and melting.

Under the current conditions the nearby gully systems are heavily modifying the lobe. Currently active seasonal flow from the surrounding gullies often disappears into the upper surface along the margin of the tongue; substantial flow also occurred within the larger stream systems close to the terminus of the tongue, leading to incision. Further field excavations during the peak of Austral summer demonstrated that a significant volume of water was migrating within the central regions of the tongue on top of the ice table, which lies at ~ 30 cm depth. Water was

observed to pond within localized depressions on the surface, indicating that the ground above the ice table had become saturated. The influx of water from gully-related surface runoff, and its lateral flow at shallow depths on the top of the ice table, appears to have reworked the soil within the active layer close to the lobe snout. The surface of this lower portion of the lobe consists of talus comprised of granite and small (5 – 50 cm) sub-angular unweathered dolerite and displays a complete lack of ventifacts, suggesting a relatively young soil age, likely younger than the formation of the lobe itself. The contrast of surface of the snout region with the upper portions of the lobe (covered with weathered dolerite blocks and boulders) suggests that the gully-related channels have a focused effect along the lobe snout.

Topographic profiles across the width of the South Fork lobate debris tongue reveal a distinct trough ~ 3 meters deep along its northern edge (Fig. 28 and 29). In contrast, the southern edge of the lobe gently slopes up to the southern flanks of South Fork. Asymmetric valley cross sections have been observed throughout the IMZ (Marchant and Head, 2007). These have been attributed to higher solar insolation along the southern (equator-facing) slopes. This insolation permitted enhanced melting and downslope movement of material through the production of meltwater, leading to the development of gentler slopes relative to the pole facing slopes (Marchant and Head, 2007). We propose that a similar environment exists along the southern slopes within SF adjacent to the lobe; this is also manifested in the form of gully erosion, which has deposited fan material that has buried the southern edge of the lobe (Fig. 28 and 29). A very similar asymmetric relationship is also observed between the gullies and the LDT in the martian study area (Fig. 28). The gully channels are larger and more pronounced along the northern (pole facing) slopes of the LDT relative to the southern slopes. The northern slope gully channels curve downslope along the edge of the LDT and have deposited fan material on the surface of the LDT in a manner that is strikingly similar to the SF lobe (Fig. 28). The reason for the 180° differences in slope asymmetry can be accounted for by the gullies in SF being dependent primarily on locations which experience above freezing temperatures (which are correlated with peak

insolation [e.g. Morgan et al., 2008]). Conversely, due to the meta stability of water under the thinner and dryer atmosphere conditions on Mars, gullies may favor sheltered slopes where surficial deposits of snow/ice could accumulate prior to melting (Hecht, 2002). Nevertheless the distinct asymmetry in slopes demonstrates the importance of insolation in controlling the modification of the lobe features on both planets.

Although additional detailed studies are necessary, the lobe in South Fork is tentatively interpreted as a relict debris-covered glacier that lost much of its original glacial ice through sublimation; after sublimation exceeded net accumulation in the accumulation zone. We propose that South Fork is analogous to the martian study site during the time period when the gullies were active on Mars. Our field investigations in the Antarctic Dry Valleys have led to the interpretations that: (1) The morphology of glacial-like features with a high debris content can be preserved following climate changes that terminate the supply of ice to the accumulation zone and render the lobes inactive. (2) Gully activity can persist in an immediately subsequent environment characterized by low precipitation and low temperatures, because the gullies can be fed by seasonal melting under peak insolation conditions of small wind-accumulated snow packs in the gully channel. (3) Prolonged gully erosion can serve to modify relict lobate features.

6. Geomorphic Interpretation of Martian Deposits and Interpretation of the Local Climate History

In this section we incorporate our field studies into our geomorphic analysis of the landforms identified in the martian study site in order to interpret the general climate conditions under which the features formed. We then use this information to broadly interpret the climatic history of the region.

The alpine glacial system in Beacon Valley (Fig. 25) provides a plausible analog for development of the Mars lobate debris tongues. The plateau ice field in Beacon Valley, which

represents the source for ice in the alpine glaciers below (Fig. 23), might be a reasonable equivalent to ice sheets believed to have been deposited by the redistribution of polar ice to lower latitudes during periods of high obliquity (Head et al., 2006a,b; Madeleine et al., 2007). The age of the LDT inferred from crater counting (surface age > 10 Ma, Fig. 19) suggests that they are significantly older than the most recent period of ice-rich mantle formation associated with the last martian 'ice age' (0.4 – 2.1 Ma, obliquity $\sim 35^\circ$; Head et al., 2003) and thus the LDT could not have formed during this time as has been suggested for other small lobate flow features (Milliken et al., 2003; Arfstom and Hartmann, 2005, see Table 1 and Fig. 31). The age of the LDT also places their formation prior to the increase in average obliquity to $\sim 35^\circ$ (Fig. 31) that occurred ~ 5 Ma (Laskar et al., 2004) that has been interpreted to explain the formation of young rock glaciers at the base of Olympus Mons (Head et al., 2005). At obliquities higher than 45° the redistribution of ice is expected to have become more pronounced and the resulting migration in ice stability (Mellon and Jakosky, 1995) may have facilitated the deposition of thick icesheets at the mid latitudes. The direct deposition of this ice onto the walls of the valleys or the calving of ice from deposits above the cliff at the summit of the slopes may have enabled the formation of small glaciers in the valleys. Debris supplied to the surface of the lobes from mass wasting along the cliffs above could have buried ice rich material and protected it from sublimation. The downslope movement of this material would have been analogous to the ice-cored lobes in Beacon Valley (Fig. 26).

Once the source of ice was cut off to the valley slopes the LDT would be dominated by sublimation which in turn is dictated by the thickness and porosity of the debris layer above the ice (Kowalewski et al., 2006). We attribute the loss of ice from the LDT as the reason for the spatial distribution of craters on the surface. If an ice core was present within the LDT, the effects of ablation of that ice and the associated deflation of the lobe would leave debris stranded on the valley floor and vulnerable to aeolian deflation, as is implied by the occurrence of elongated ridges along the southern valley. The exposure of boulders and other debris during the deflation process

would generate a rough surface that would be a poor recorder of impact craters, hence the distinct lack of craters along the snout of the LDT in Figure 19. Further ablation of the LDT and a loss of ice from the ice core could cause subsidence of the surface and generate the elongated depressions that are observed on the surface which are similar to ‘spoon depressions’ observed in terrestrial debris rich glaciers that have endured enhanced ablation. The lack of craters on these depressions (Fig. 19) may indicate that the sublimation is on going, and therefore, ice has been preserved within the LDT since their formation prior to 10 Ma. This level of ice preservation beneath a debris layer is consistent with geomorphic (Head et al, 2006a,b) and ground probing radar (Holt et al., 2008; Plaut et al., 2009) data that support the occurrence of large bodies of glacial ice below a relatively thin sublimation lag within LDA/LVF deposits which are estimated to be 100 – 500 Ma (e.g. Mangold, 2003; Morgan et al., 2009a, see Fig. 31).

The occurrence of a similar texture and features interpreted to be related to the occurrence of near surface ice (e.g. boulders with moats and elongated depressions, see Fig. 16) along the floor of valleys < 5 km wide to those present on the surface of the LDT suggests that ice rich deposits may also be preserved within these valleys. Such deposits could have been formed the same way as the LDT, except that the lack of a steep gradient prevented the formation of obvious flow features (e.g. compressional ridges). The absence of these features within the wider valleys is logical as the floors of thinner valleys would be more sheltered from insolation and would receive higher amounts of debris due to the closer proximity of the valley walls, and therefore would provide a higher preservation potential for ice.

The occurrence of elongated ridges along the outer edge of a degraded LDT (Fig. 17) argues for a relationship between the two features. The ridges are morphologically similar to terrestrial glacial moraines and thus might represent material and sediment deposited by former LDT prior to their retreat or total deflation as a result of sublimation since their initial formation. The location of the ridges along the base of south facing valley walls supports this interpretation as equator-facing slopes will experience higher temperatures due to favorable solar insolation and

thus ice rich lobate features would be more prone to sublimation along these slopes than slopes with other orientations. Additionally, the better preserved LDT shown in Figure 14 all are within heads of valleys and thus are flanked on three sides by slopes which can provide a source of debris. In contrast the southern valley walls extend in a relatively straight direction for >40 km and thus ice-rich deposits formed along them would have been supplied debris from a smaller unit length of cliff face. Therefore, we interpret the ridges to be indicative of the demise of LDT as a result of climatic change. The reason for the difference in degradation state between the different LDT deposits is not interpreted to be because of climatic effects but is due to the lower preservation potential of ice rich deposits formed along exposed equator facing slopes.

The modeling work of Morgan et al (2009b) suggests that the gullies within the study area were active during periods of high obliquity throughout recent martian history. Though the level of activity is interpreted to be dependent on the aspect of the gullies, with the most recent activity limited to the pole facing slopes and occurred as recently as ~0.5 Ma. These age estimates are consistent with other estimates of gully activity (Reiss et al., 2004; Schon et al., 2009, see table 1) and demonstrates the significant difference in age between the gullies and the LDT. Fieldwork conducted in the ADV has shown that gully activity can occur due to the melt of small scale aeolian-deposited snow banks. Therefore, gully activity requires substantially less ice to be deposited than is required for the LDT formation, and unless there were prolonged periods of low obliquity, gullies are likely to have been forming since the initiation of the LDT. The aspect dependence of gully morphology has also had an apparent affect on the gradient of the slopes on the valleys (Fig 2c). The gentler slopes are pole facing and are correlated with the largest and most complex gullies, which appear to have assisted the downslope movement of material. The increase in erosion on polefacing slopes might have been too affective for LDT preservation and may explain the absence of LDT or ridges at the bases of slopes of this orientation (Fig. 22).

The occurrence of a potentially relict gully channel below a LDT is intriguing (Fig. 17, 18). One possibility is that the channel relates to a period of gully activity that occurred prior to

the initiation of the LDT. Statistical analysis of the evolution of Mars' obliquity has shown 37.62° to be the average value for the last 5 Ga (Laskar et al., 2004). This value is consistent with gully formation models (e.g. Costard et al., 2002) and demonstrates that gully activity may have occurred throughout the Amazonian. If this is correct the formation of LDT might represent a limited deviation in obliquity to higher values that could enable the deposition of large volumes of ice in the mid latitudes. Alternatively the relict channel may be related to the LDT lobe, and therefore might be representative of the melting of ice within the LDT. Milliken et al (2003) suggested that VFF may have provided a source of water to gullies elsewhere on Mars. No other antiquated channels were found in proximity to LDT termini. Therefore, if melting had occurred it does not appear to have been widespread. Nevertheless, the occurrence of small fluvial channels in association with lobate features or gully forms that appear stratigraphically older than typical Malin and Edgett (2000a) gullies should be a target for future surveys using high resolution data sets.

7. Conclusions

Our research underlines the potential for ice-related landforms on Mars to be used to derive information about climatic changes that occurred during the Amazonian period of the history of Mars. The study site has experienced a long and diverse geological history, involving the formation of Asimov crater in the Noachian, its subsequent infill by sedimentary material and its capping by thick lava flows. Although the process of formation of the valleys is still a matter of debate, the valleys themselves have produced a unique environment in which to study the geomorphic signals of climate change. The occurrence of steep slopes ($>20^\circ$), relatively narrow (hence sheltered) valleys, and a source of debris has provided favorable conditions for the preservation of ice and the formation of gullies within preferential insolation environments. Analysis of the study site has led to the identification of a distinctive class of features; lobate

debris tongues (LDT). These are small-scale, ice and debris-rich flow features, and their emplacement largely predates the formation of gullies. Ridges along the floors of the valleys are interpreted to indicate the former extent of LDT and related ice-rich lobes, and thus are further indication of climatic changes that have affected the region.

Detailed mapping in the ADV yields terrestrial analogs that provide insight into the origin of LDT in Noachis Terra (i.e., upper Beacon Valley; Fig. 25), their late-stage modification, and subsequent gully incision (i.e., present-day South Fork, Upper Wright Valley; Figs. 27-30). The observed morphologic changes are best explained as a shift in martian climate from one compatible with excess snowfall, accumulation and glacial flow, to one incompatible with excess snowfall and glaciation, but consistent with minor snowmelt and gully formation. Available dating suggests that the climate transition occurred >10 Ma, prior to the formation of other small scale ice rich flow features identified elsewhere on Mars that have been interpreted to have formed during the most recent phases of high obliquity. This suggests that multiple climatic shifts may have occurred over the last tens of millions of years of martian history.

Acknowledgements

Special thanks are extended to my advisor **James W. Head** and to **David R. Marchant** for extensive discussions on this work during the formulation of the ideas and content of this chapter. I look forward to additional assistance and collaboration with the two of them as this contribution is prepared for publication.

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Tables.

Landform	Location/Description	Age Estimate	Rationale	Reference
Gullies	Gully fans superimposed on a dune field within Nirgal Vallis at 29°S, 318°E.	0.3 – 1.4 Ma	Based on crater counts of the dune field.	<i>Reiss et al (2004)</i>
Gullies	Gully with fan superimposed with secondaries at ~35°S, 131°E.	1.25 Ma	Based on crater count age date of the primary crater that produced secondaries	<i>Schon et al (2009)</i>
Gullies	Study Area: Gullies display strong aspect dependent morphology	Pole facing gullies >0.5 Ma Equator facing gullies >5 Ma	Based on correlation with Laskar et al (2004) simulations of obliquity variations, suggests pole facing gullies >0.5 Ma, equator facing gullies >5 Ma	<i>Morgan et al (2009b)</i>
Lobate Flow Features	VFF. Mid-latitudes in both hemispheres.	0.1 – 30 Ma Though stratigraphic association with latitude dependent mantle suggests a similar age: 0.4 – 1.25 Ma	Based on average strain rates for viscous flow within a 10 m thick dust-rich icy layer under martian conditions	<i>Milliken et al (2003)</i>
Lobate Flow Features	GLF (presumably of similar origin to VFF documented in Milliken et al., 2003) in crater at 40°S, 67°E.	~ < 10 Ma	Crater counts	<i>Arfstrom and Hartmann (2005)</i>
Lobate Flow Features	‘Rock glaciers’ along the NW edge of Olympus Mons scarp	~ > 5 Ma	Crater counts	<i>Neukum et al (2004), Head et al (2005)</i>
Lobate Flow Features	LDT in study region	> 10 Ma	Crater counts	<i>This paper</i>
Lobate Flow Features	LDA/LVF. Northern dichotomy boundary	100 – 500 Ma.	Crater counts	<i>e.g. Mangold (2003), Morgan et al (2009a)</i>

Table 1.

Table 1. Age estimates of gullies and ice-rich flow features on Mars reported in the literature. Observations of distinct fan stratigraphy suggests that multiple events of gully activity have occurred (e.g. Schon et al., 2009), thus the above ages should be considered a date of the most recent activity. Note that for crater counts on small surface areas (such as the surface of the smaller ice-rich features: Arfstrom and Hartmann., 2005; LDT), the uncertainties surrounding the production function of the smallest diameter craters, the effect of secondary craters and the nature of erosion and deposition, the inferred ages represent order-of-magnitude estimates (Arfstrom and Hartmann., 2005).

Figure Captions

Fig. 1. Regional setting of the study area on Mars. MOLA shaded relief overlain on MOLA gridded data.

Fig. 2. The study area within the degraded Asimov crater in Noachis Terra located at 46°S, 5°E. (a) The interior of the crater contains an annulus of valley systems and a off-center depression. CTX data over HRSC image: h1932_0000. (b) DTM of the study region. The degraded rim of the Asimov can be seen in the topographic data to the northwest. The deepest portion of valleys is 2000 m below the average surface elevation of the crater floor. The dashed white lines are the location of topographic transects in Fig. 4. DTM derived from HRSC orbit: h1932_0000. (c) Slope map of slopes >15° in the study area. Slope measurements derived from HRSC DTM orbit: h1932_0000.

Fig. 3. Location of figures and names of valleys within the study area. CTX data over HRSC image: h1932_0000

Fig. 4. Topographic profiles across Asimov crater. The interior of the crater has a relatively consistent elevation of ~500 m, the floors of the valleys and the central depression are 300-2000 m below this elevation, suggesting that the crater is filled with 2 km of material. Topography derived from HRSC DTM orbit: h1932_0000.

Fig. 5. Exposed columnar jointing along the edge of the rock layer that caps the valley walls. (a) Polygonal fractures 10 m in diameter are present within the surface of the rock layer close to the top of the valley slopes. (b) Perspective view of 30m high structural columns within the rock layer. The columns have the same diameter as the polygonal fractures seen in (a). This structure is similar to columnar jointing that forms in terrestrial basaltic flows as a result of thermal contraction during cooling. This suggests that the rock layer was formed by a lava flow on top of the crater fill. HiRISE images: PSP_003603_1325 (a), PSP_003669_1325 (b).

Fig. 6. Evidence for mass wasting in the form of slumping along the slopes of the valleys. Boulders within the slump blocks suggest that the blocks originated from the rock layer at the top of the slopes. Arcuate fractures along the surfaces above the valley slopes imply that extension has occurred, which may contribute to further detachment of the valley slopes. (a) Western slope of the northeastern valley complex. HiRISE image: PSP_003814_1325. (b) Northern slope within the central depression. CTX image: P05_003102_1327. See Fig. 6 for location.

Fig. 7 . Morphology of the latitude dependent mantle within the study area. (a) Along the top of the valley slopes the mantle forms a distinct layer above the coherent lava flow rock layer. The abrupt change in slope along the edge of the rock layer provides a perspective view of the mantle, which in some areas appears to be several tens of meters thick and consist of multiple layers. In some places, the occurrence of boulders emerging from the layers suggests that these

are regions where the mantle is relatively thin (near the cliff edge) and the structure of the underlying rock layers shows through. (b) Example of degradational textures along the surface of the mantle, including isolated pits and shallow cusate depressions. These features have been attributed to the loss of ice through sublimation (Mustard et al, 2001). HiRISE images: PSP_003880_1325 (a) PSP_003603_1325 (b).

Fig. 8. Mosaic of THEMIS infrared nighttime images covering the study area. The relatively high temperatures along the top of the valleys (bright pixels) are correlated with a discrete rock layer (see Fig. 5) and are thus likely caused by the high thermal inertia of the rock layer (and associated boulders) relative to loose sediments and dust that cover the rest of the surface.

Fig. 9. Aspect dependence of the morphology of gullies within the study site. (a) Gullies on pole-facing slopes are highly incised and display dendritic tributaries that merge into single ~100 m wide channels. (b) Gullies on equator-facing slopes originate from cusate, ~500 m wide alcoves cut into the upper portion of the slopes. Multiple, relatively straight, 50 m wide channels emerge from the apex of the alcoves. (c) High resolution image of pole-facing gully channels that display fluvial features, including small ~5 m-wide channels that are present within the larger channel systems. (d) Gullies on west-facing slopes are also well incised, but are not as densely eroded into the slopes and do not display tributaries. (a) and (b): CTX: P07_003880_1327. (c) and (d) HiRISE image: PSP_004091_1325.

Fig. 10. Isolated Gullies. (a) Gullies eroded into slopes on either side of an isolated ridge along the southern end of the northeastern valley; boxes represent the location of the adjacent images. (b) The occurrence of gullies on an isolated slope here and elsewhere within the study area, argues against a groundwater source for the water that carved the gullies. (c) Zoomed in

view of gullies in (b), demonstrating that they do not feel the effect of the ridge. Morgan et al. (2009b) suggested that the gullies were formed by the melting of snow deposits. The image in (b) also shows the aspect dependence of gully morphology present throughout the valley systems. a): HRSC: h1932_0000 b): HiRISE: PSP_003880_1325.

Fig. 11. Small gullies along a debris covered mound on the floor of the southern valley. The gullies display the three morphologic components of other martain gullies: alcove, channels and fan (Malin and Edgett, 2000a). The small size of these features suggests they are geologically very young, though they might have been active at the same time as the larger gully systems along the upper slopes of the valley walls. HiRISE image: PSP_006926_1320.

Fig. 12. Small-scale gullies along sand dunes. These gully systems are morphologically different from gullies elsewhere in the study area, although they are similar (despite being an order of magnitude smaller) than gullies observed on Russell Crater dunes (right image). Left image: HiRISE: PSP_006926_1320. Right image: HiRISE: PSP_007018_1255

Fig. 13. Phase diagram in temperature and pressure coordinates for water, showing the annual climatic conditions between the top of the valley walls (where the standard gullies originate) and the lower flanks (where the smaller scale gullies originate). Average surface pressures and maximum surface temperatures are plotted for the course of a Martian year in each case. The temperatures are the same due to the small martian atmospheric lapse rate, but the elevation difference does have an effect on surface pressures. The plots show that conditions above the triple point are only met for a short period each day during the spring at the elevation of the smaller gullies, and even less for the alcove elevation of the standard gullies. Surface pressures derived from the work of Lobitz et al (2001). Surface temperatures predicted by the 1D LMD/GCM model used by Costard et al. (2002).

Fig. 14. Examples of lobate debris tongues (LDT) within the study area. White arrows indicate the termini of the lobes, the large black arrows indicate the inferred direction of flow. (a) 4 km long LDT within the southern end of the northeastern valley system. Gullies are present along all of the slopes that surround the LDT, the distal ends of the gully channels and their fans curve down along the surface of the lobe. The small black arrows highlight small indentations along the edge of the LDT, which may have formed as the result of strain induced by brittle deformation. HiRISE image: PSP_003880_1325 (b) 3 km long LDT that is being fed by smaller tributaries within the western portion of the valley. CTX image: P07_003880_1325 (c) Broad LDT along the eastern end of the southern valley. Compressional ridges can be seen on the surface of the LDT (d). CTX image: P07_003880_1327 (c), HiRISE image: PSP_004091_1325 (d).

Fig. 15. Topographic transect of LDT along the eastern edge of the southern valley (a), and accompanying MOLA profile (b). The topographical profile was derived from extracting adjacent MOLA points from different orbital tracks. HiRISE image: P06_003313_1327.

Fig. 16. Features interpreted to be ice-related, along the surface of LDT and the floors of the adjacent valleys. (a) Large boulders (>5m) on the surface of the LDT have moats surrounding them; these may be formed due to the loss of near surface ice, via sublimation as a result of the thermal properties of the rocks. (b) Elongated troughs, 200 m wide, that are parallel to the long axis of the LDT. The troughs may have been formed by the lowering of the LDT surface as a result of the loss of subsurface ice through sublimation. (c) Similar trough to (b) along the floor of the valley to the north of the LDT. (d) Further to the north, concentric troughs form 50 m wide stepped terraces. Boulders with moats are also present along the floor of the valley. The occurrence of these features on LDT and the floor of the valleys suggest that both features contain

near surface ice and presumably formed the same way via the accumulation of ice and debris.

Context image and (d): CTX image: P07_003880_1327. Other images, subsets of HiRISE image: PSP_003880_1325 .

Fig. 17. Image and accompanying MOLA profile of a portion of the southern wall of the southern valley that shows a degraded LDT superimposed on a relict gully channel. Gullies are also superposed on the surface of the LDT, opening the possibility that the formation of the LDT predated a previous period of gully formation. Composite of CTX images. The box represents the location of the image in Fig. 18.

Fig. 18. Image (a) and accompanying sketch map (b) showing a potential relict gully channel that is stratigraphically beneath the now-degraded LDT. This suggests that at least some gully activity was occurring prior to the formation of the LDT. Elongated ridges are also present along the surface of the lobe. MOC image: M1101304.

Fig. 19. Distribution of craters along the surface of a LDT. There is a distinct absence of craters along the lobe snout and along the depressions in the center, suggesting that the surface of the LDT has undergone modification, possibly as a result of ablation. HiRISE image: PSP_003880_1325.

Fig. 20. Crater size-frequency diagram for the craters mapped on the surface of a LDT (Fig. 19). Red diamonds represent the actual counts of craters on the surface. The blue diamonds represent the predicted crater size distribution once the effect of the steep valley walls filtering out oblique impacts is considered. This suggests an age for the LDT of >10 Ma. Isochrons plotted according to Hartmann (2005).

Fig. 21. Schematic figure to demonstrate the effect steep valley walls will have on filtering out oblique impacts for a point exactly half way between the two valley walls. For the LDT counted in Fig. 19, θ was 35° . Note, not to scale

Fig. 22. Location of the ridges within the southern valley (highlighted with black lines). Note that the ridges are found along the base of the north-facing slope. The location of images in Fig. 23 are bounded by the boxes. Composite of CTX and HRSC images.

Fig. 23. High-resolution views of the elongated ridges (highlighted by white arrows). (a) Near continuous ridge of hummocky terrain that extends for ~ 5 km along the base of the western end of the southern valley. CTX image: P15_006715_1327. (b) Example of ridges displaying sharp crests along the surface of the ridge. MOC image: S1800854. (c) Example of arcuate ridges at the base of the slope. CTX image: P05_003102_1327.

Fig. 24. Satellite view of the Antarctic Dry Valleys. Fieldwork was conducted in the SUZ along the northwestern wall of Beacon Valley (highlighted by blue box) and in the most elevated portion of the IMZ in South Fork of Upper Wright Valley (highlighted by red box). Image: Landsat 7/ETM+. Taken on December 18, 1999.

Fig. 25. Alpine glaciers and ice cored lobes in Beacon Valley, Antarctica. (a) perspective view of LIDAR DEM. (b) shows a photograph taken in the field. The alpine glacier acquires ice and snow from avalanches along the edge of the plateau icefield. Well-developed ice-cored moraines are visible in both data sets; under the influence of gravity these flow downslope, forming ice-cored lobes. The DTM demonstrates the topographic similarities between Beacon Valley and the martian study area. The development of ice-cored lobes may provide a terrestrial

analog for the formation of LDT. LIDAR data: NSF/NASA/USGS with processing by Schenck et al. (2004).

Fig. 26. Close up photographs of the ice-cored lobes in Beacon Valley. (a) The lighting in the top image highlights the steep lobe fronts. (b) Perspective from the snout of the lobe in the white box in (a). The surface of the lobes consists of unconsolidated loose angular boulders. For reference the largest boulder in the center of the frame is ~ 1 m long.

Fig. 27. Ground temperature recorded at top of gully fan in SF between summer 2006 and summer 2008 (blue line) compared to equivalent seasons for a martian southern hemisphere (47° S), pole-facing slope at 35° obliquity (red line). Pink represents period of melting conditions ($T > 273$ K) on both planets, blue represents periods of surface ice stability ($T < 200$ K) for Mars, i.e. when accumulated snow and ice would be stable. Mars model generated by 1D version of LMD/GCM (as used by Costard et al., 2002).

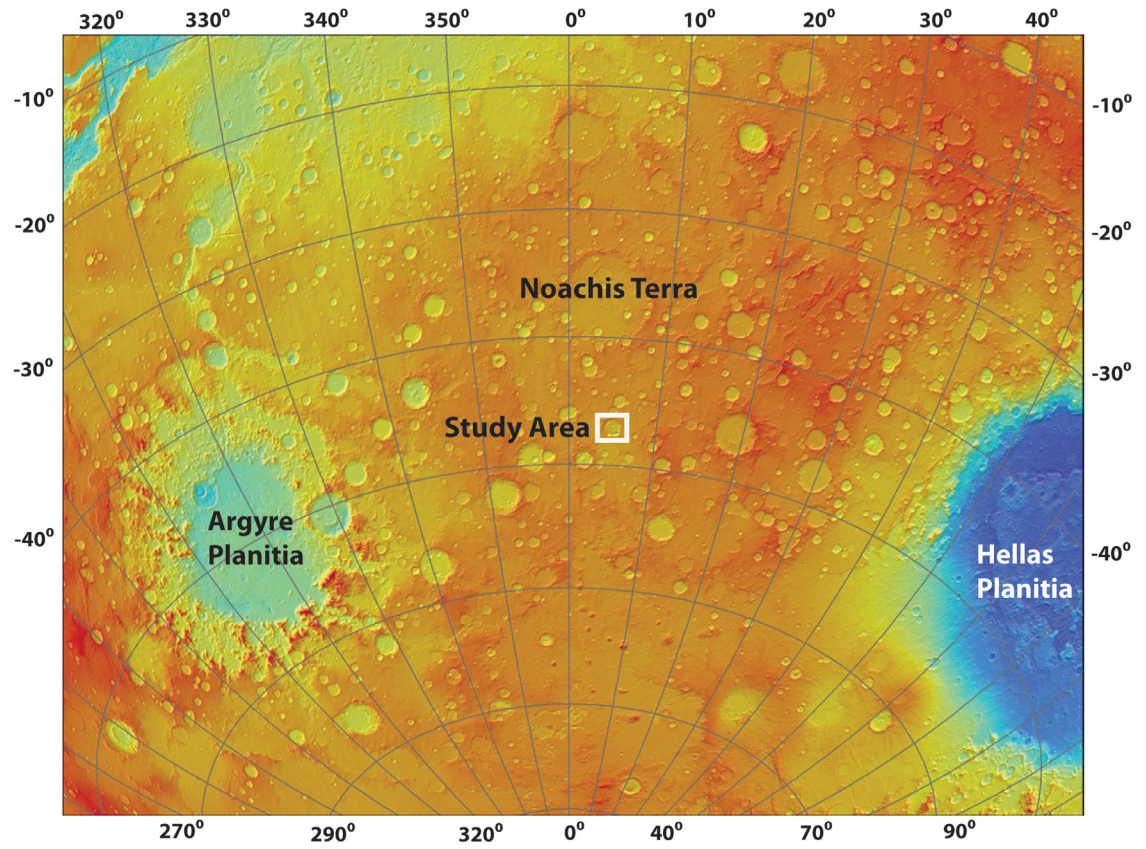
Fig. 28. Morphological comparison between a martian LDT (a) and the lobe in South Fork, Antarctica (b). Both features display convex-up topographic profiles, debris covered surfaces and are surrounded by gullied slopes. Our research suggests that both features are relict and have been modified by gullies since they were rendered inactive by climate change. Arrows indicate North. (a) HiRISE image: PSP_003880_1325. (b) Ikonos image: po105577

Fig. 29. Topography of the South Fork lobe. A sharp contact can be seen between the lobe and the northern wall of South Fork. In contrast, the southern wall, which is eroded by gullies, appears to be burying the edge of the lobe. Slope asymmetry is observed throughout the IMZ, and is related to the availability of liquid water on equator-facing (warmer) slopes (Marchant and Head., 2007). The white dashed lines represent topographic profiles in Fig. 30.

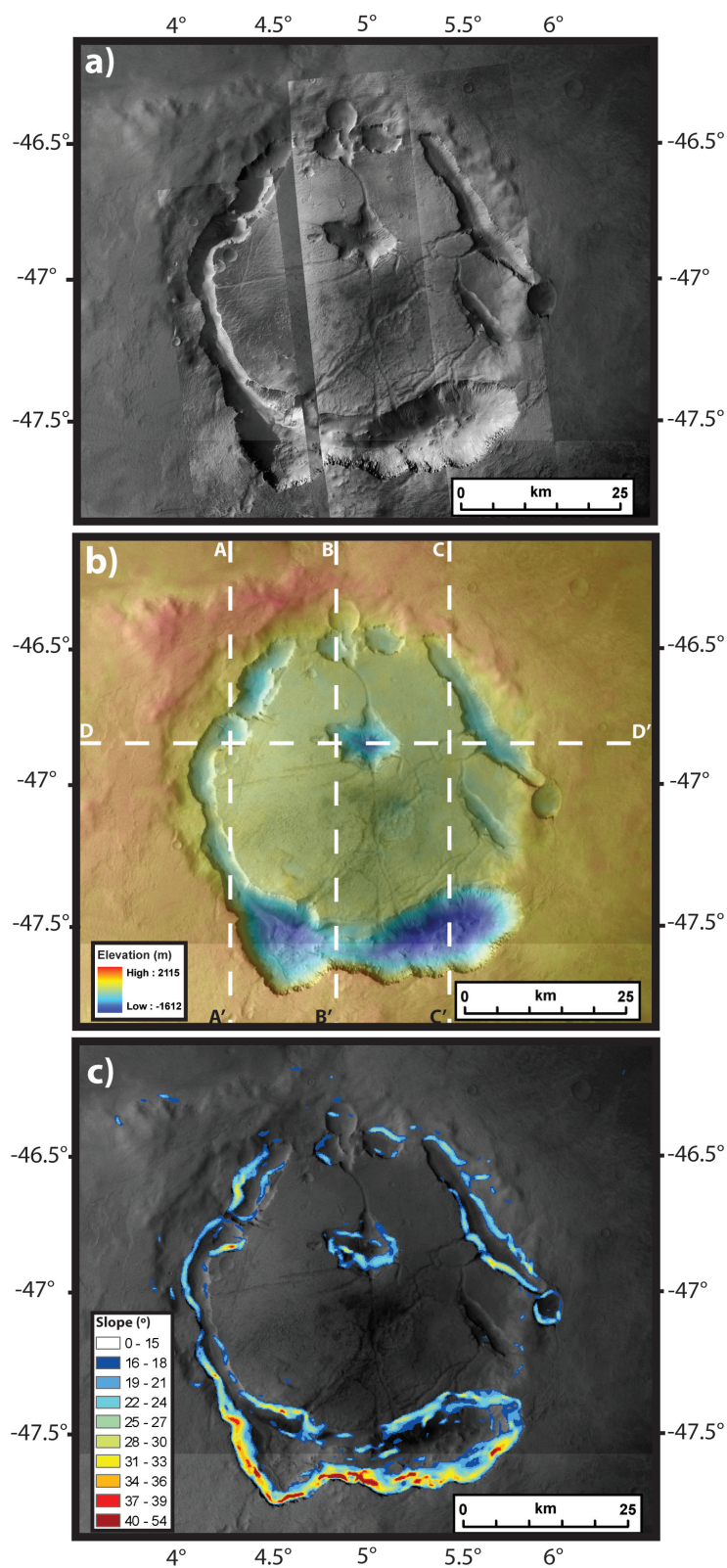
LIDAR topography overlain LIDAR hillshade. LIDAR data: NSF/NASA/USGS with processing by Schenck et al. (2004).

Fig. 30. Topographic profiles of the South Fork lobe. The south-north profiles (A-A' to C-C') reveal the valley slope asymmetry, with gentler slopes along the gullied north-facing valley walls. Similar slope profiles have been observed throughout the IMZ and have been attributed to downslope movement of material due to meltwater generation along the warmer north-facing slopes (Marchant and Head, 2007). In South Fork this has caused the burial of the southern edge of the lobe. Gully channels eroded along the surface of the lobe itself are visible in the profiles. The west to east facing transect demonstrates the convex-up nature of the snout of the lobe which is similar to the terminus of the LDT in Fig. 15.

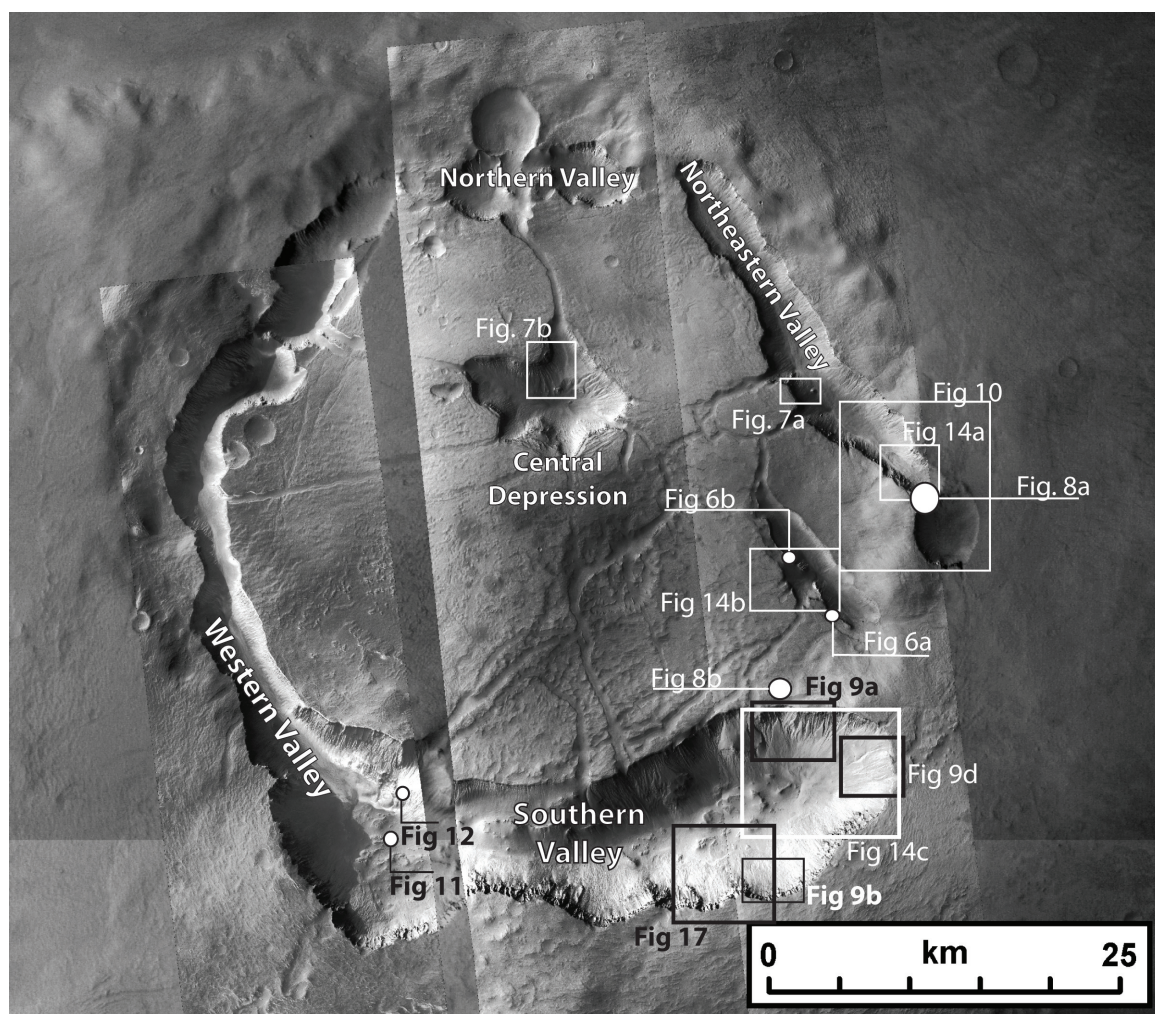
Fig. 31. Comparisons between the age of the LDT and the age of other lobate viscous flow features on Mars (See Table 1 for more details). Crater counts of the surface of the LDT (Fig. 19, 20) reveal that it has an age >10 Ma. This is prior to the most recent period of high obliquity that has been interpreted to have deposited the latitude dependant mantle (LDM) during an 'ice age' (Head et al., 2003) and formed Viscous Flow Feature (VFF) (Milliken et al., 2003) and Glacier Like Feature (GLF) (Arfstrom and Hartmann, 2005). The LDT also formed prior to the transition from higher obliquity (average $\sim 35^\circ$) to lower obliquity (average value $\sim 25^\circ$). The plot also demonstrates the significantly younger age of the LDT relative to LVF and LDA deposits, suggesting that large-scale glacial deposition has not occurred over the most recent period of the Late Amazonian. Obliquity plot from Laskar et al (2004). Obliquity becomes non-deterministic prior to ~ 20 Ma; nevertheless statistical assessment of the possible values demonstrates that the average value of obliquity over the last 5 Ga was likely to be $\sim 37.6^\circ$.



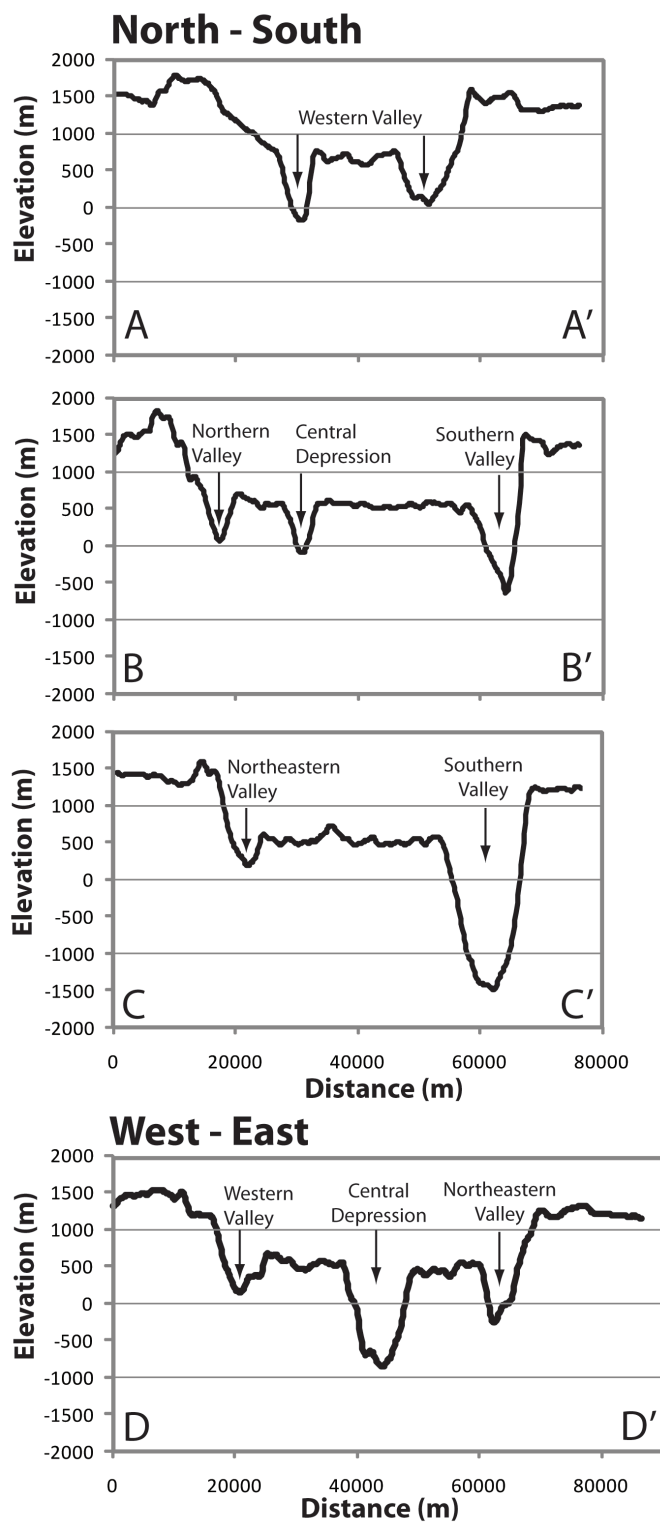
Chapter 6, Figure. 1



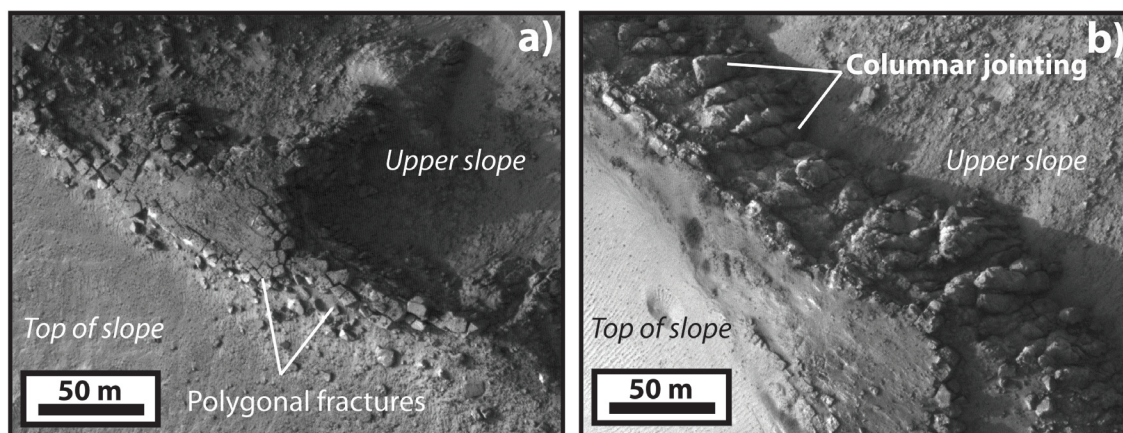
Chapter 6, Figure. 2



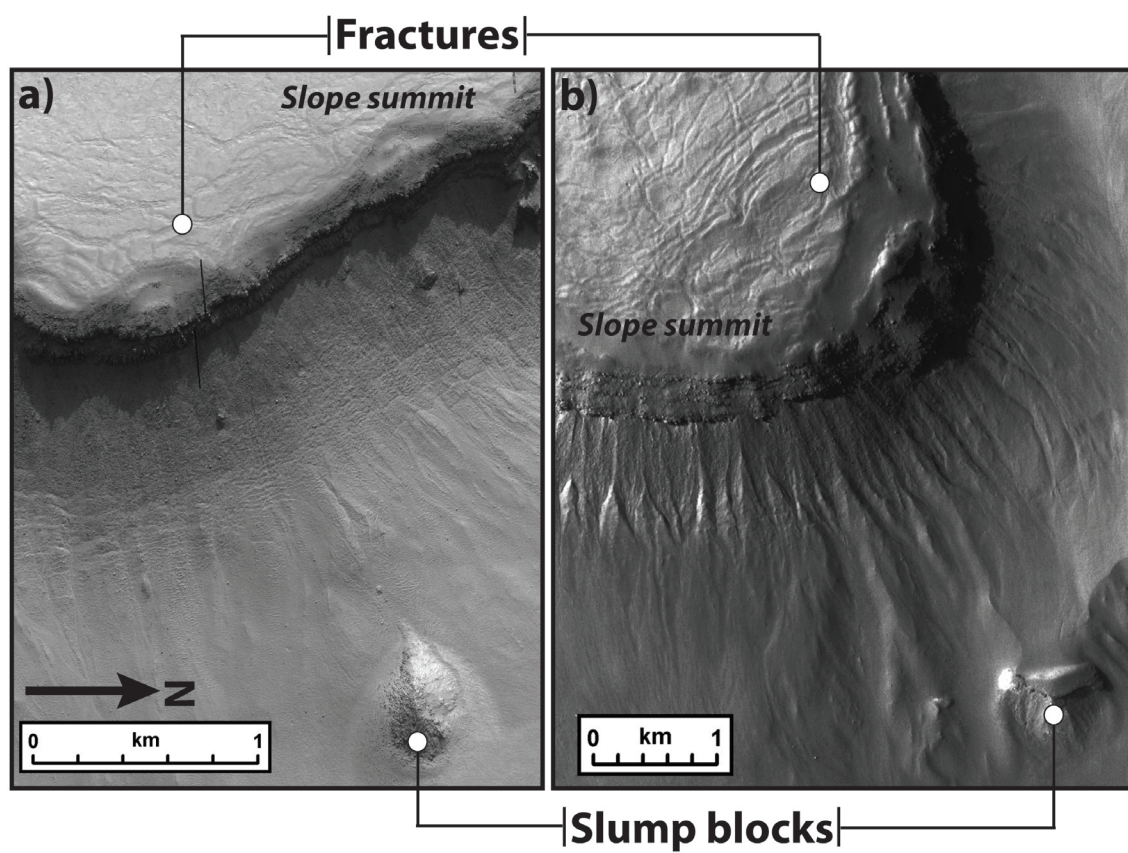
Chapter 6, Figure. 3



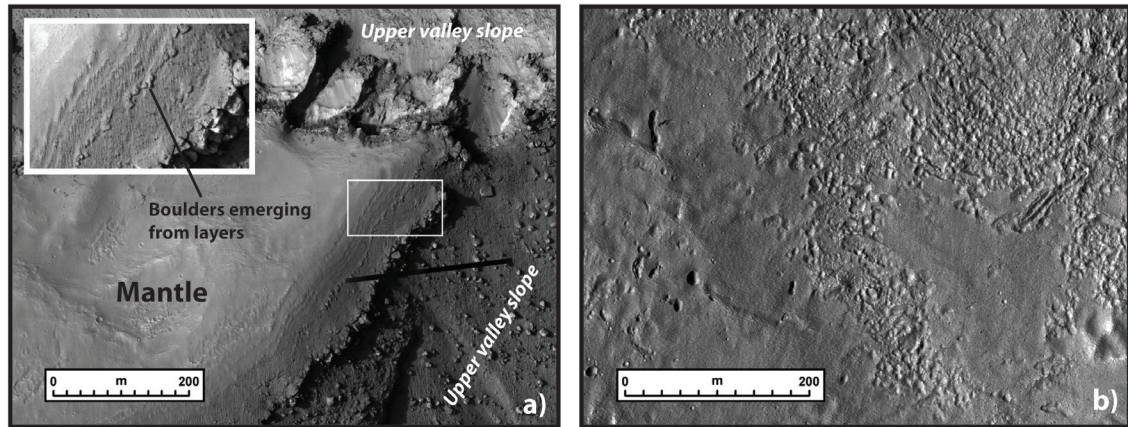
Chapter 6, Figure. 4



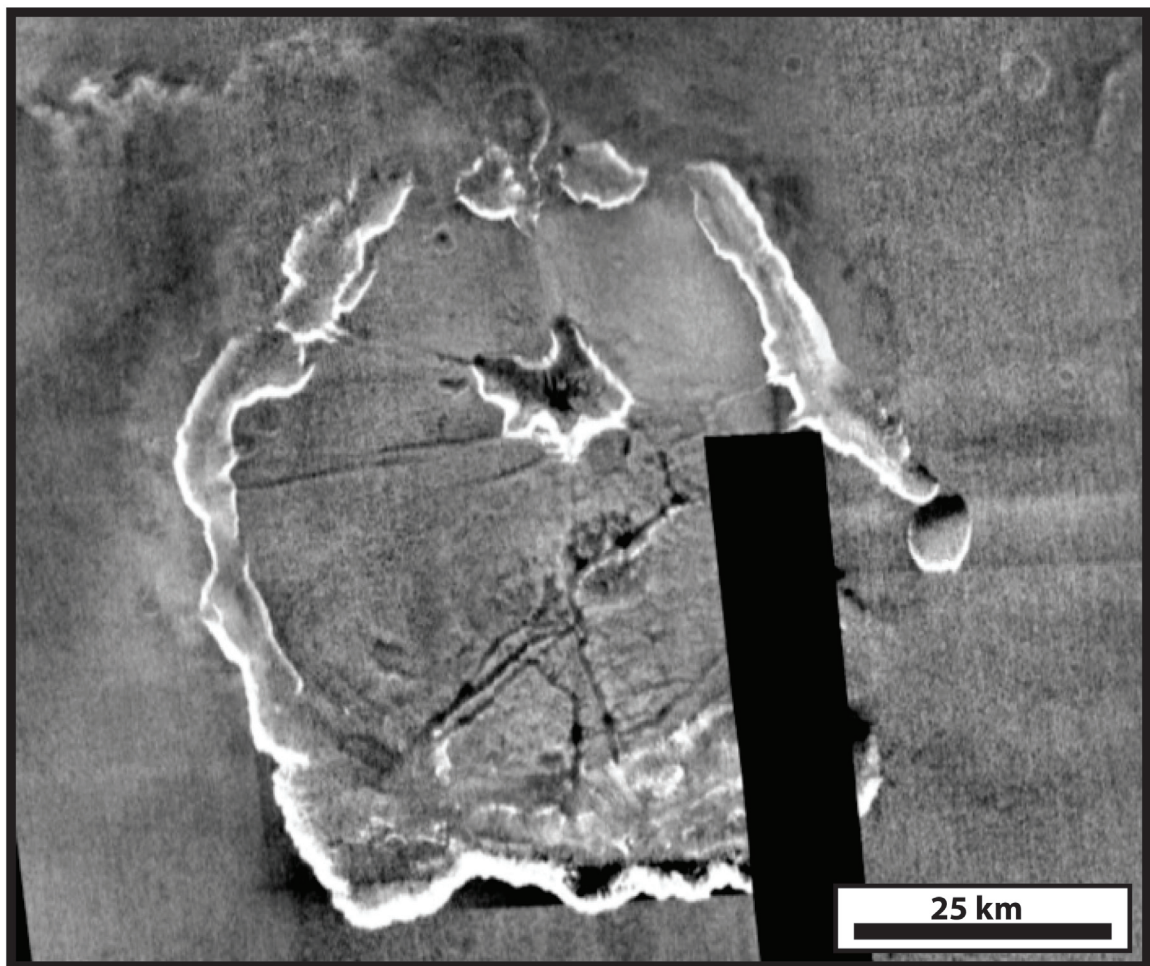
Chapter 6, Figure. 5



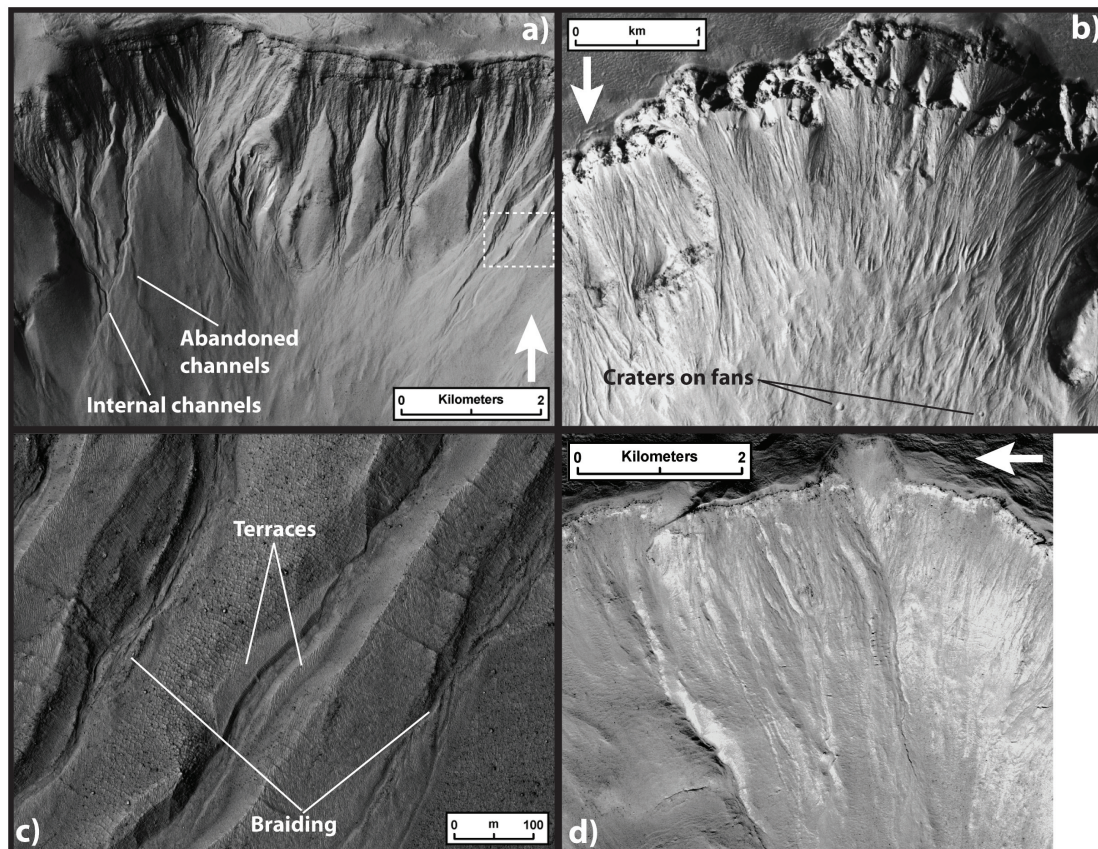
Chapter 6, Figure. 6



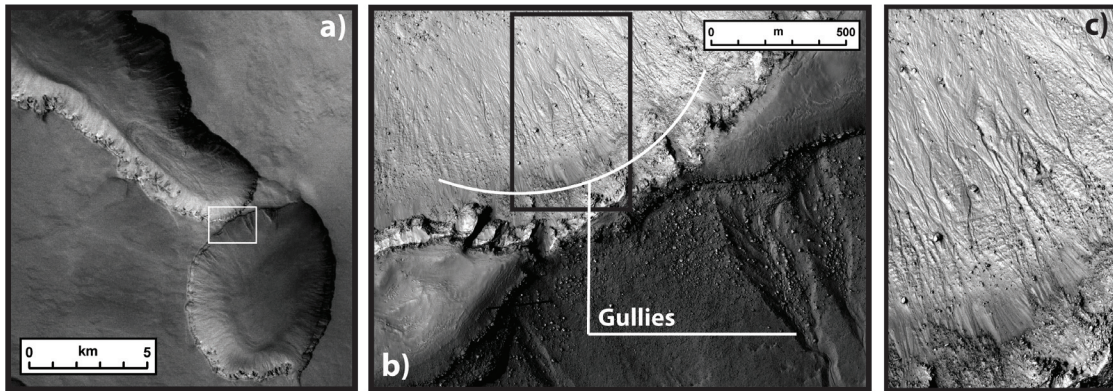
Chapter 6, Figure. 7



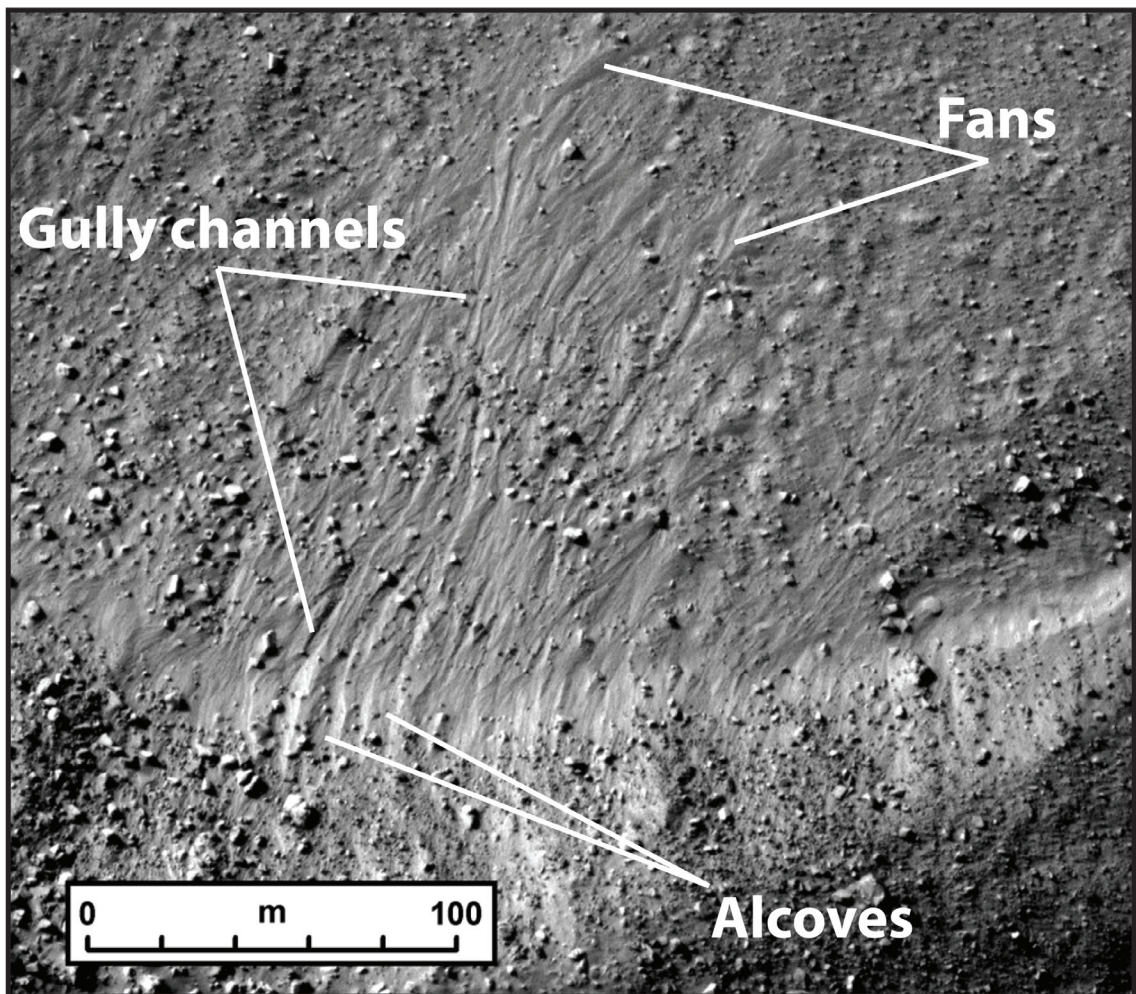
Chapter 6, Figure. 8



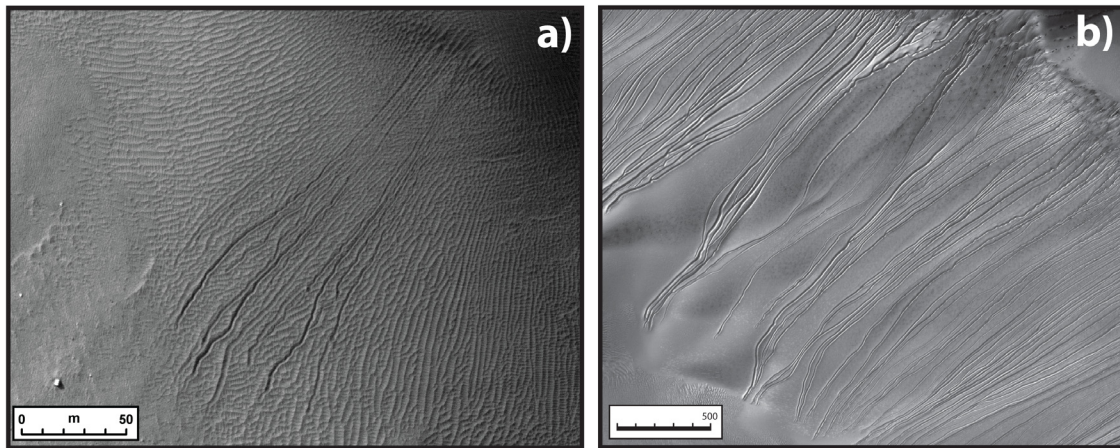
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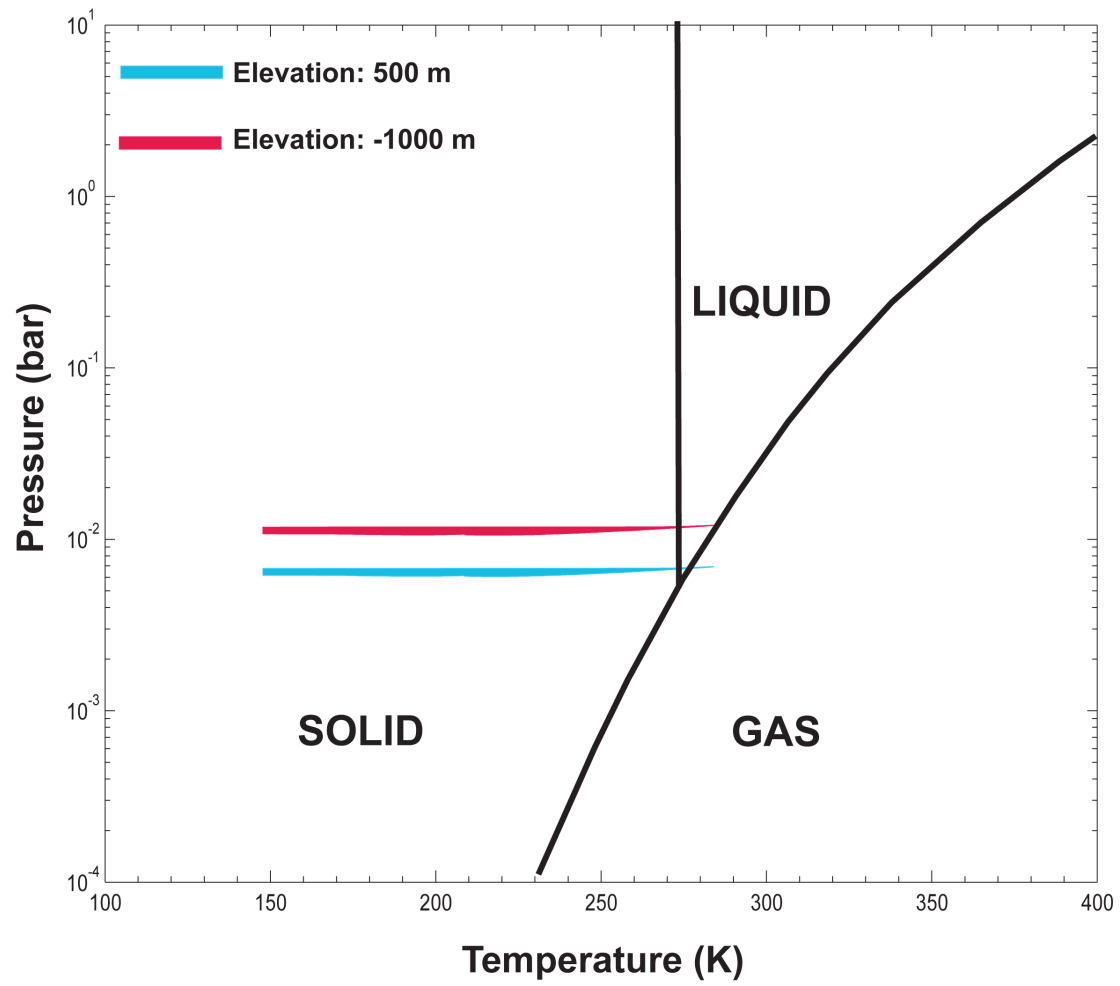
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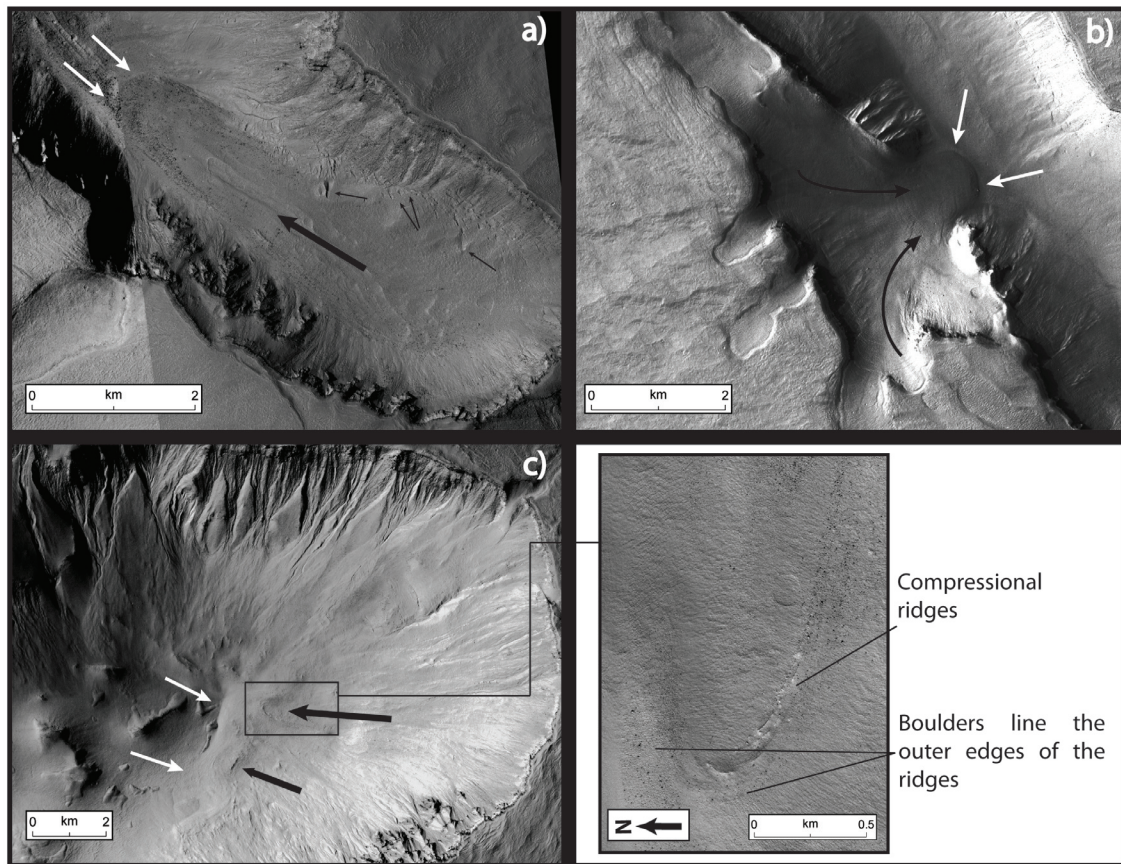
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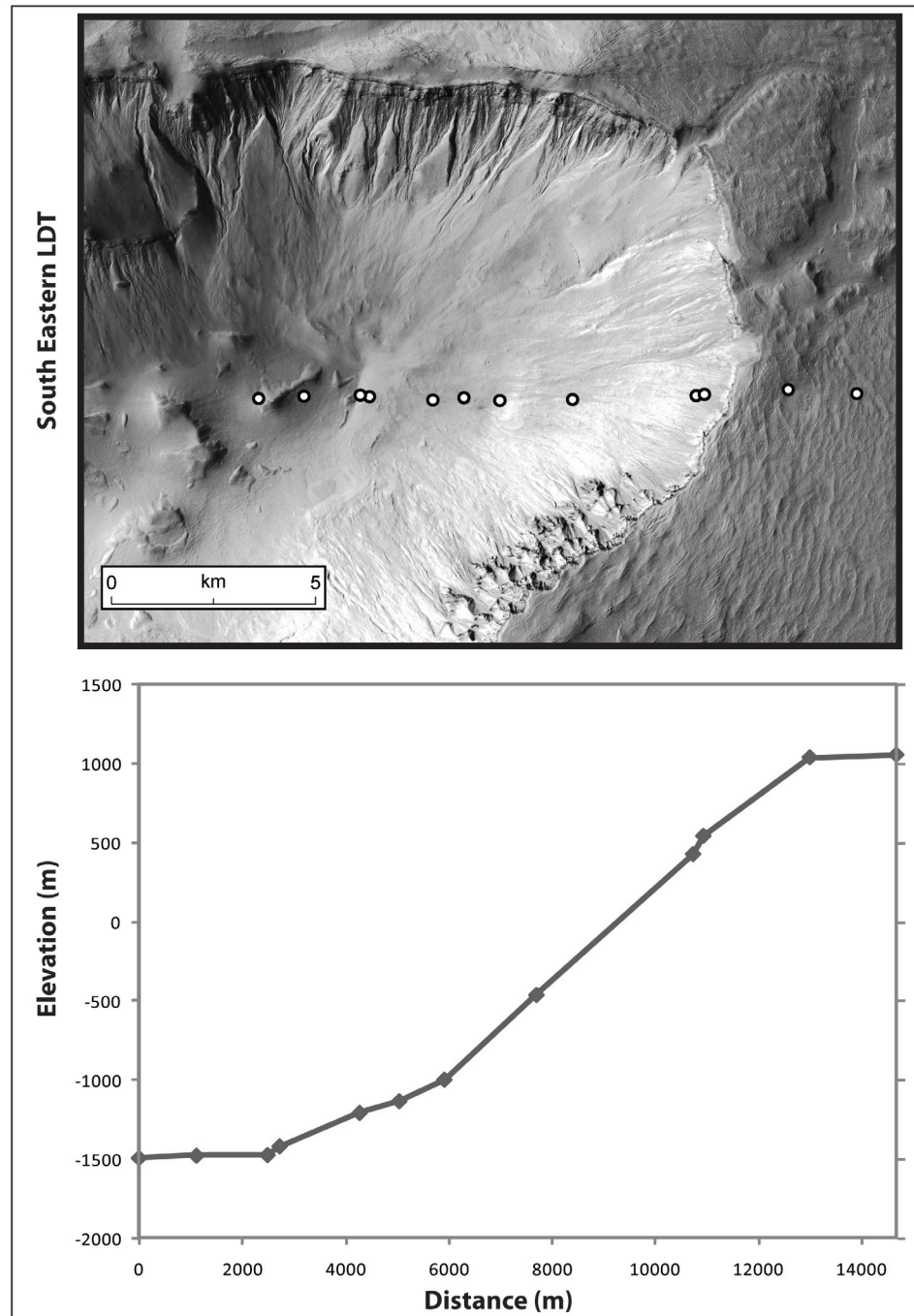
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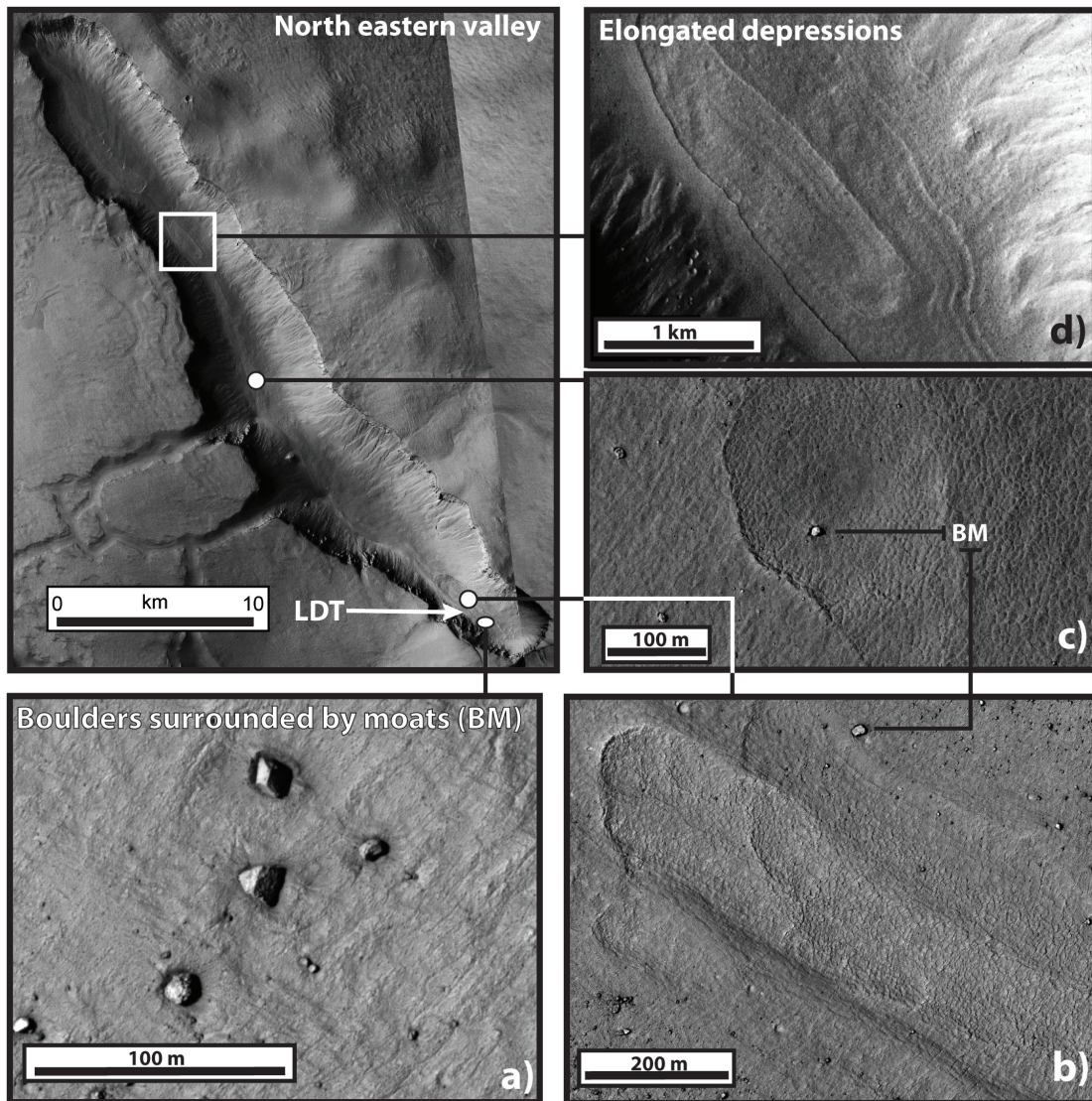
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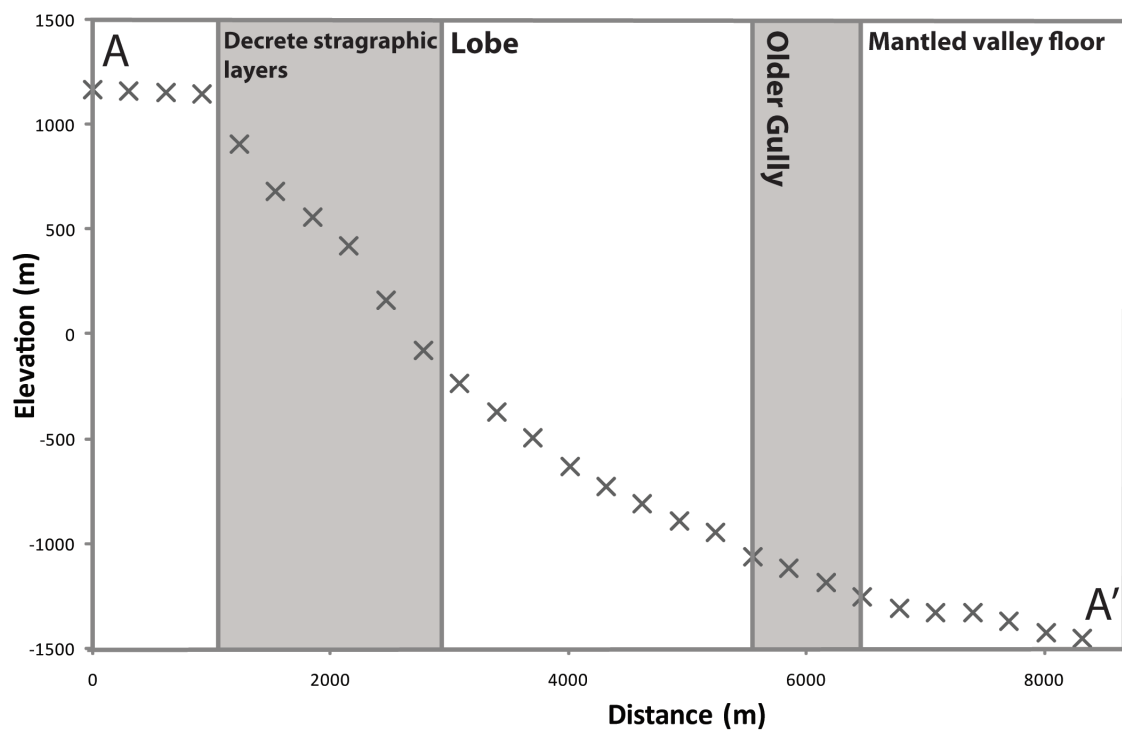
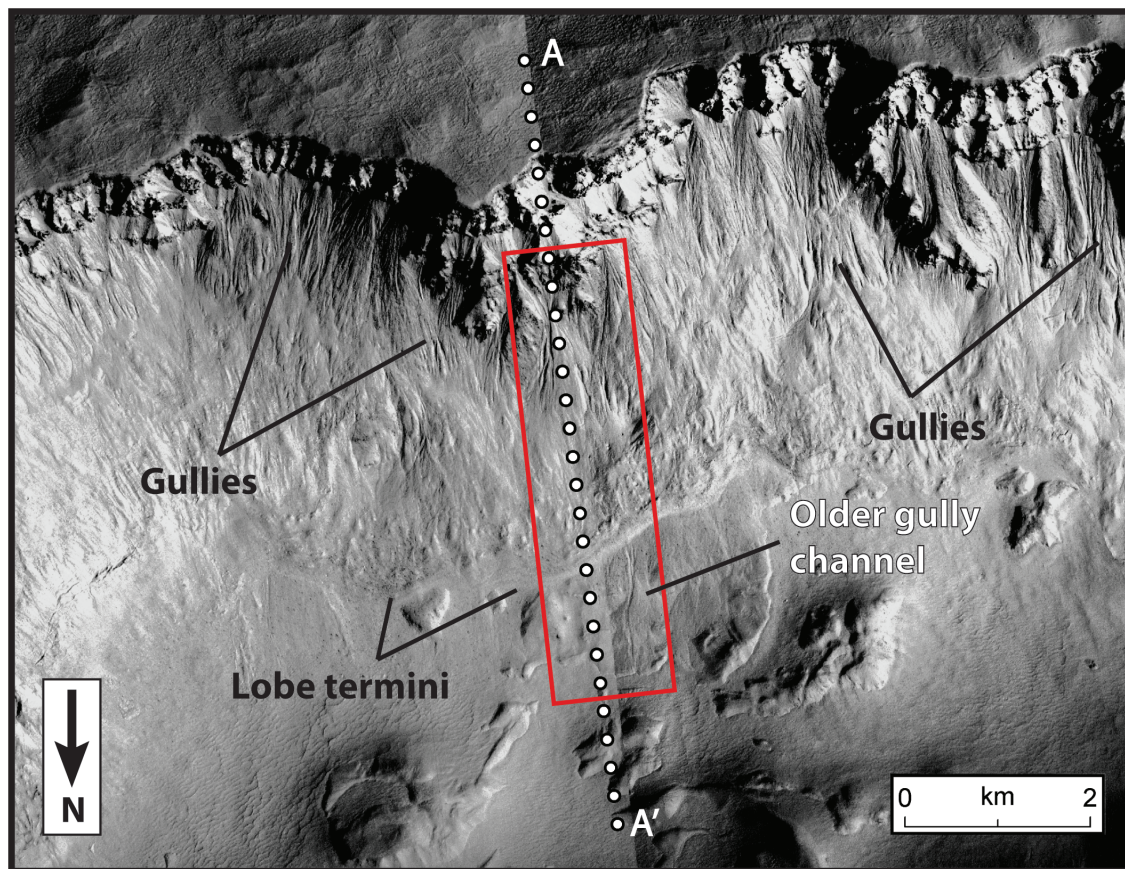
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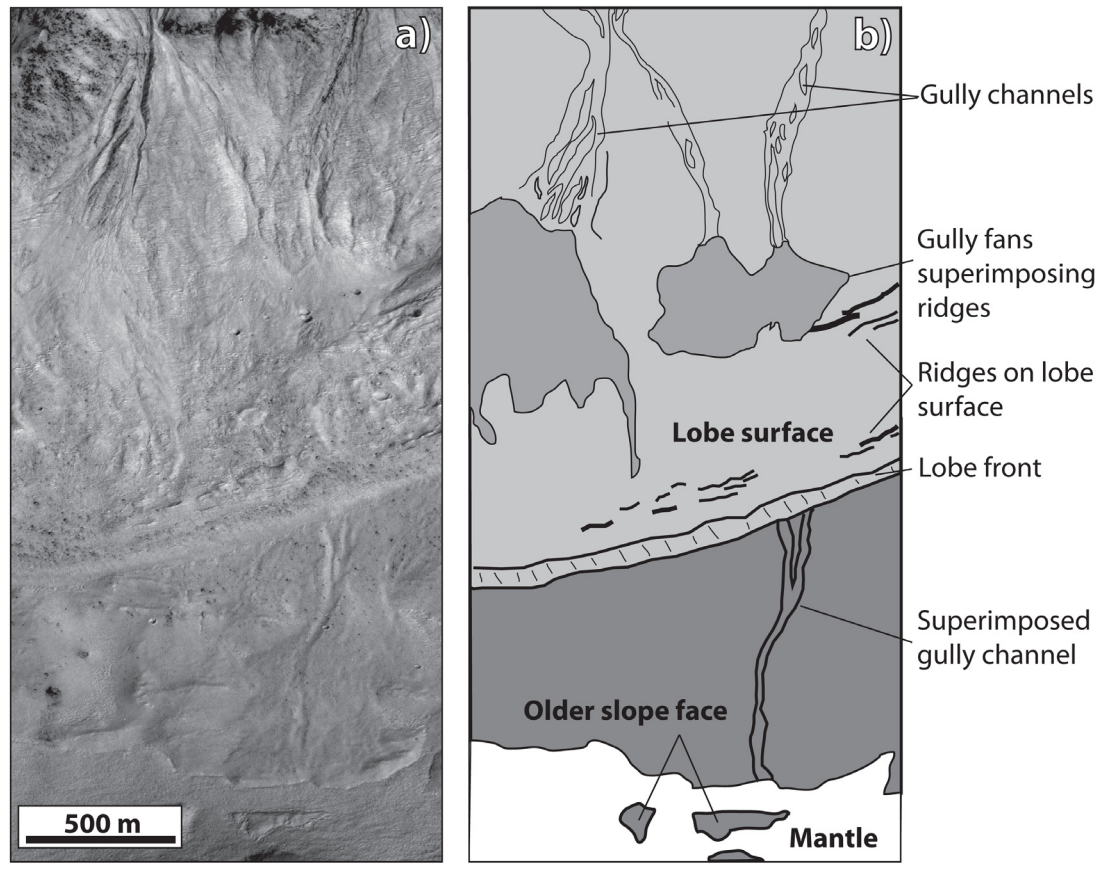
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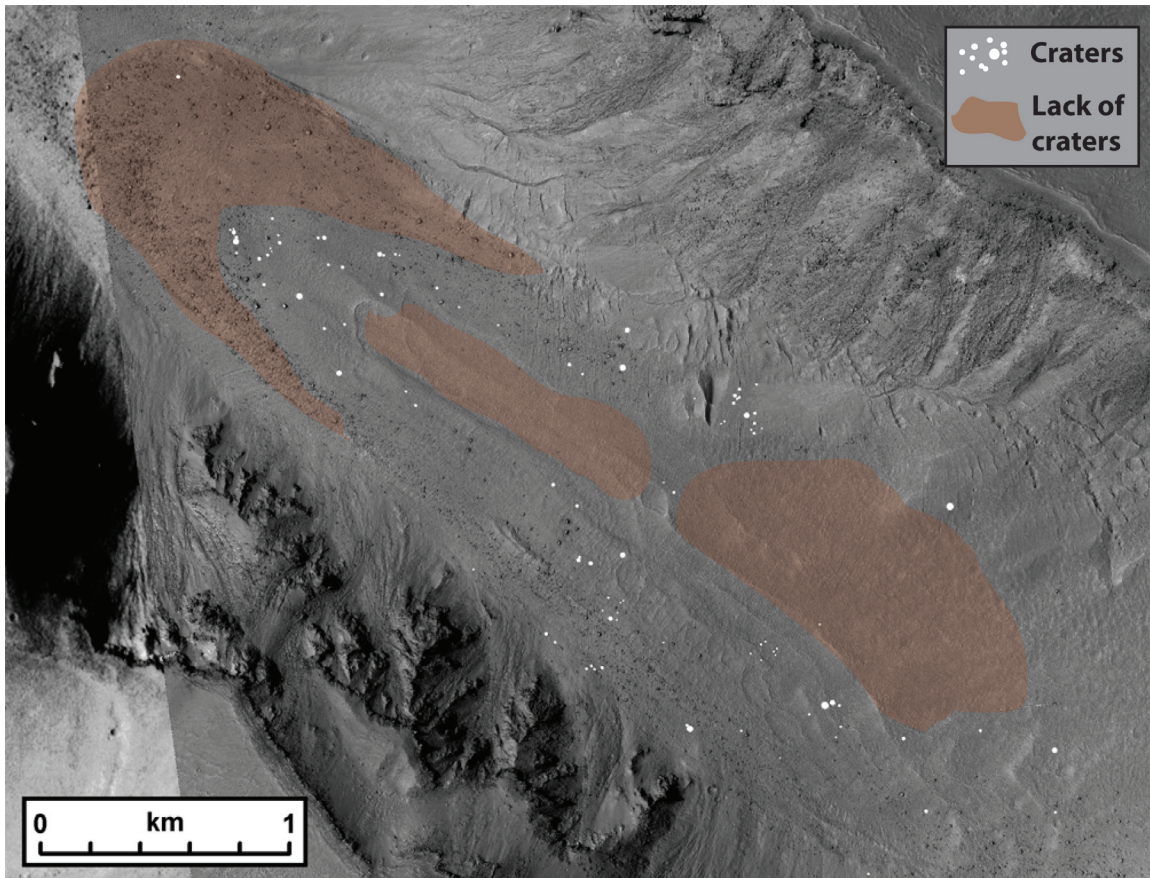
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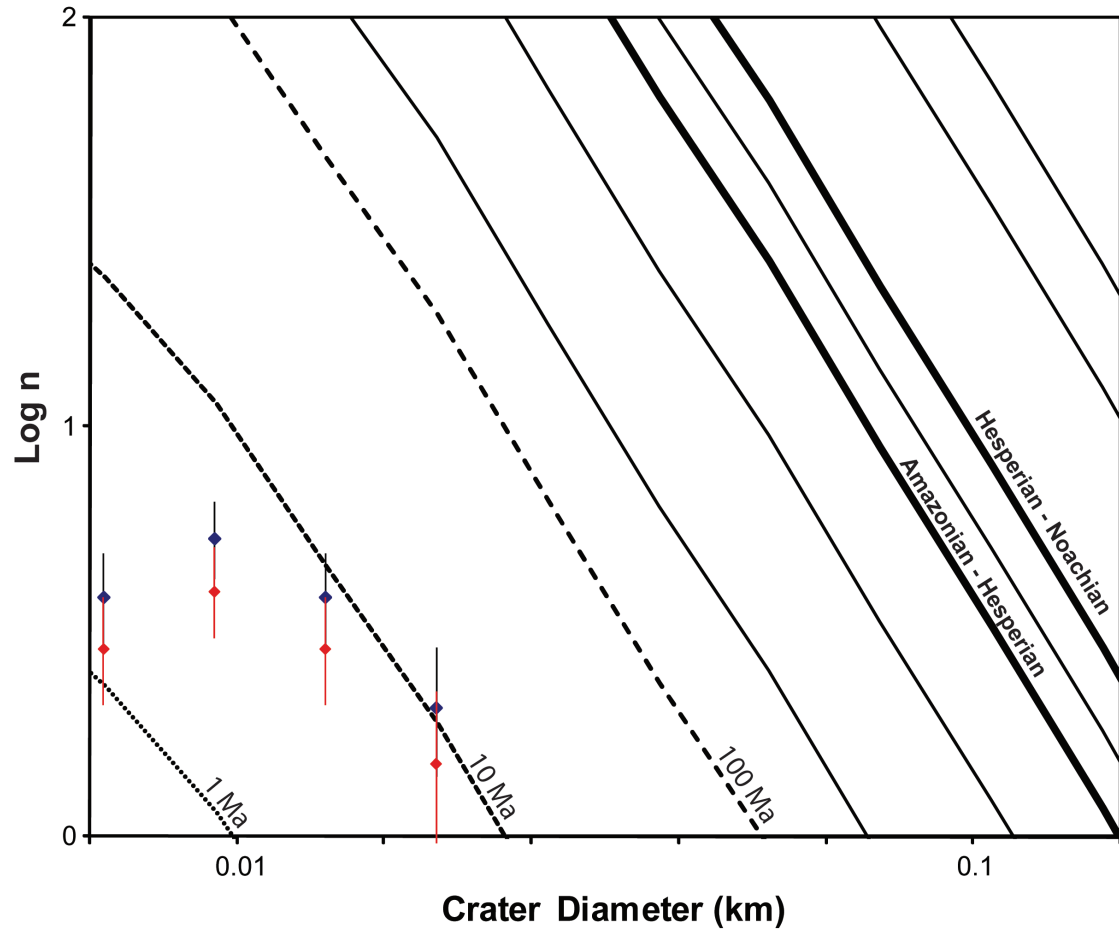
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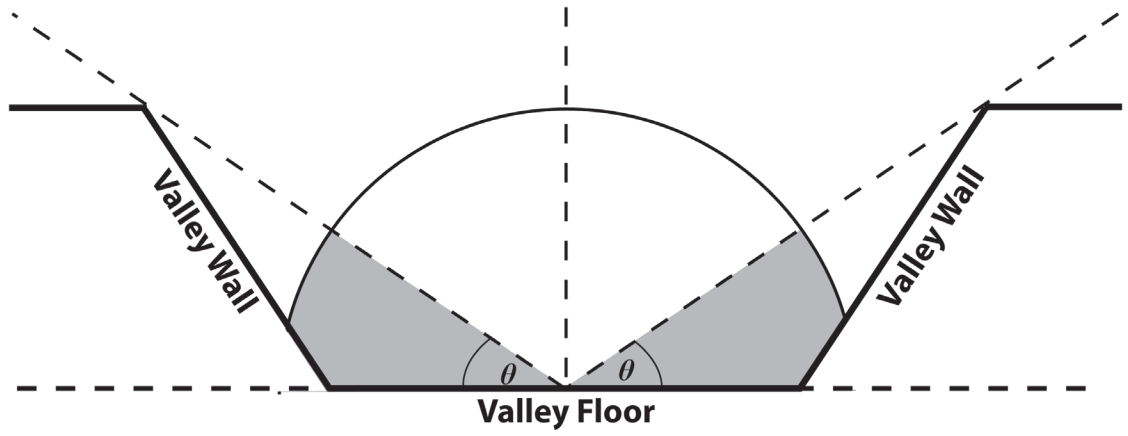
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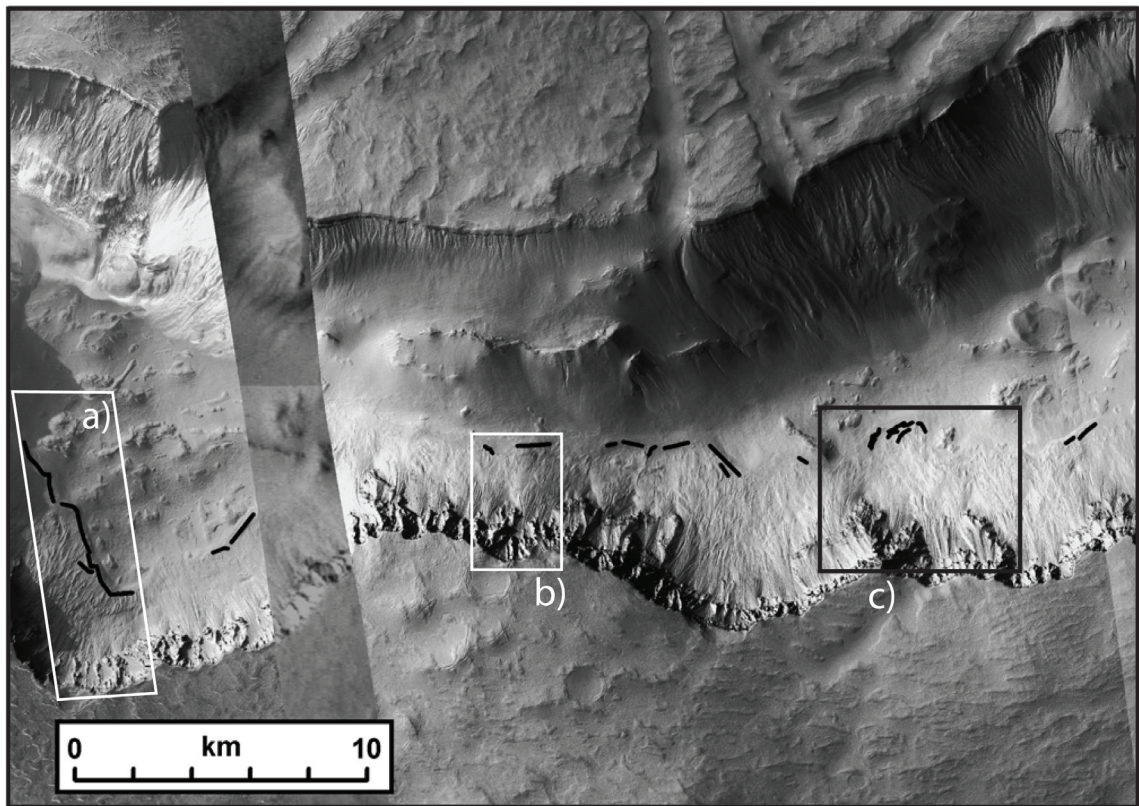
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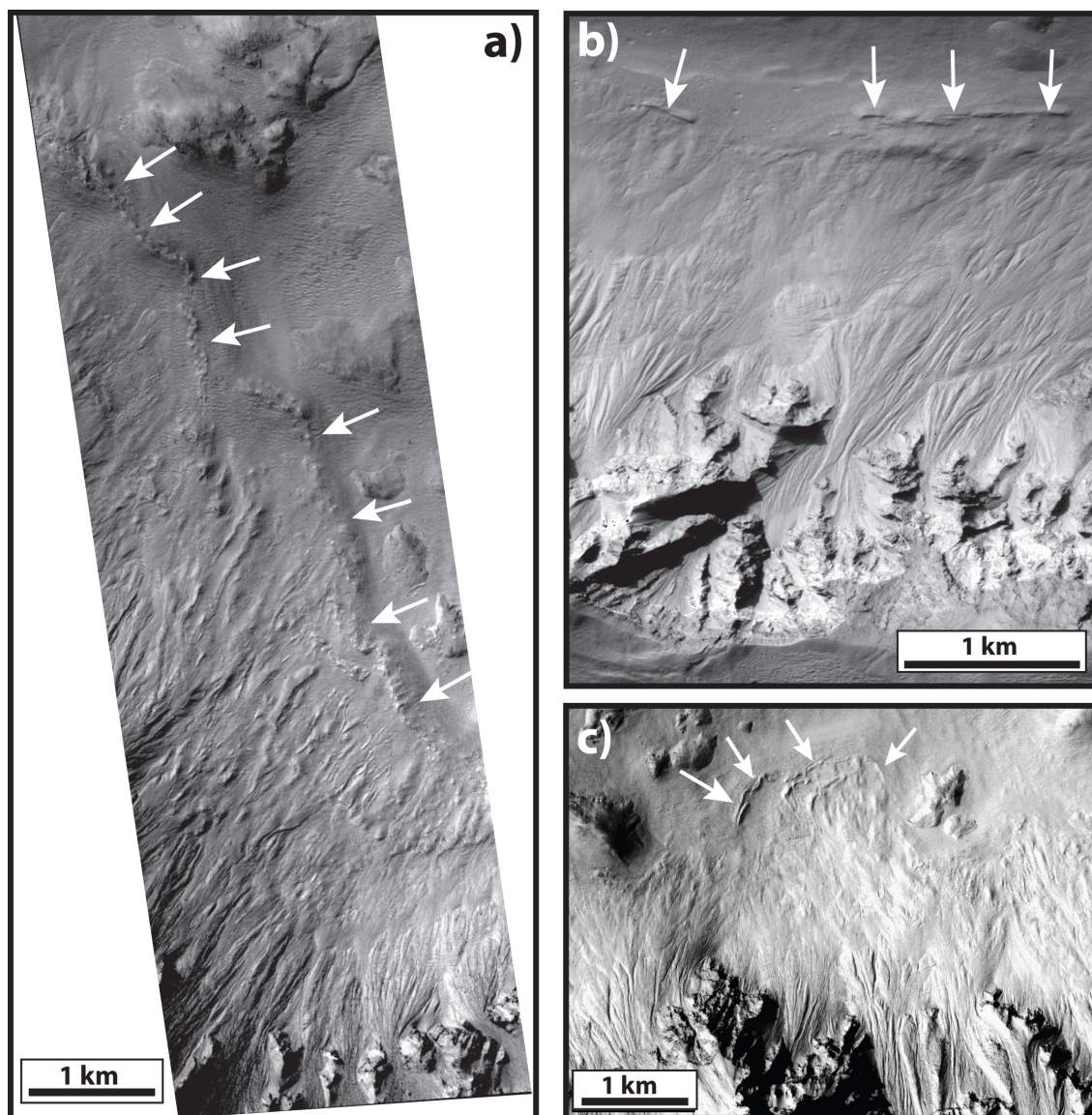
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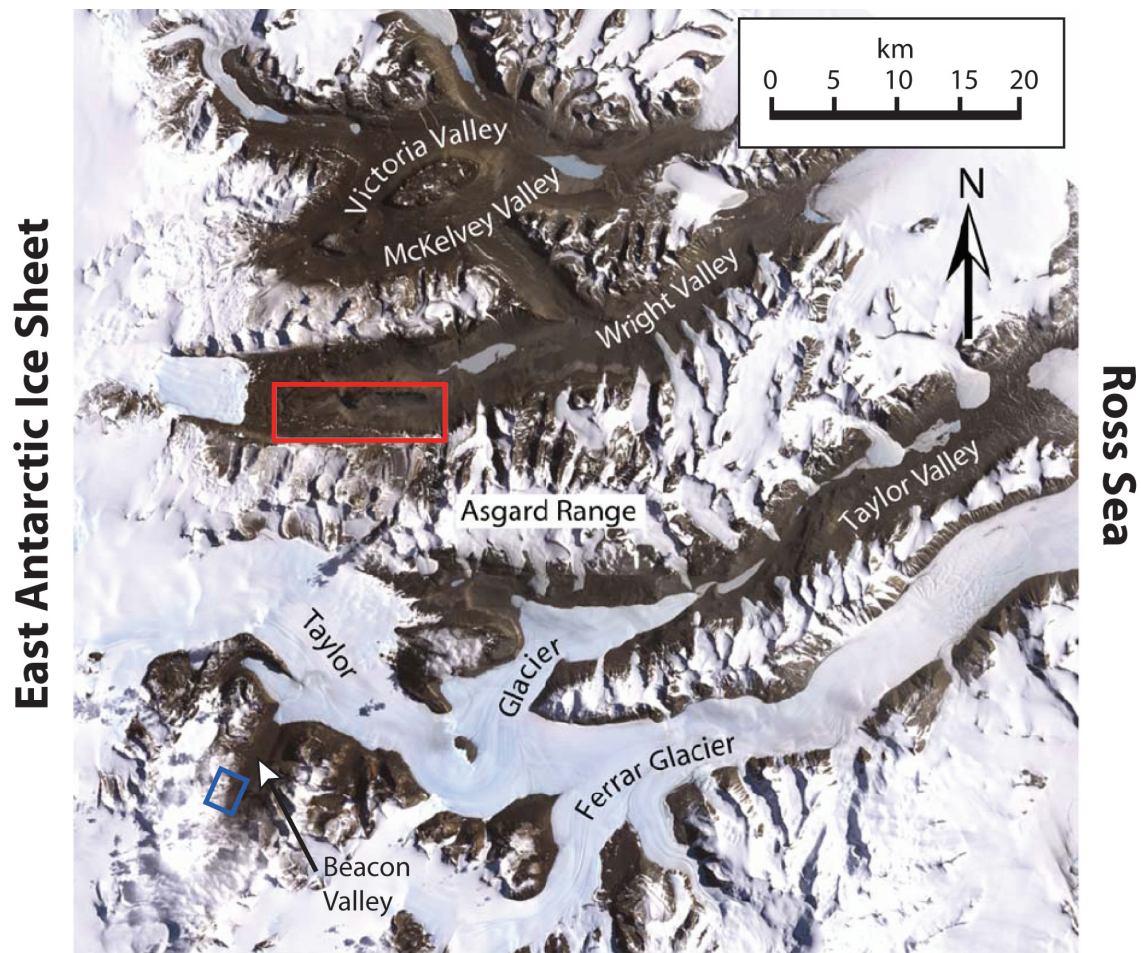
Chapter 6, Figure. 21



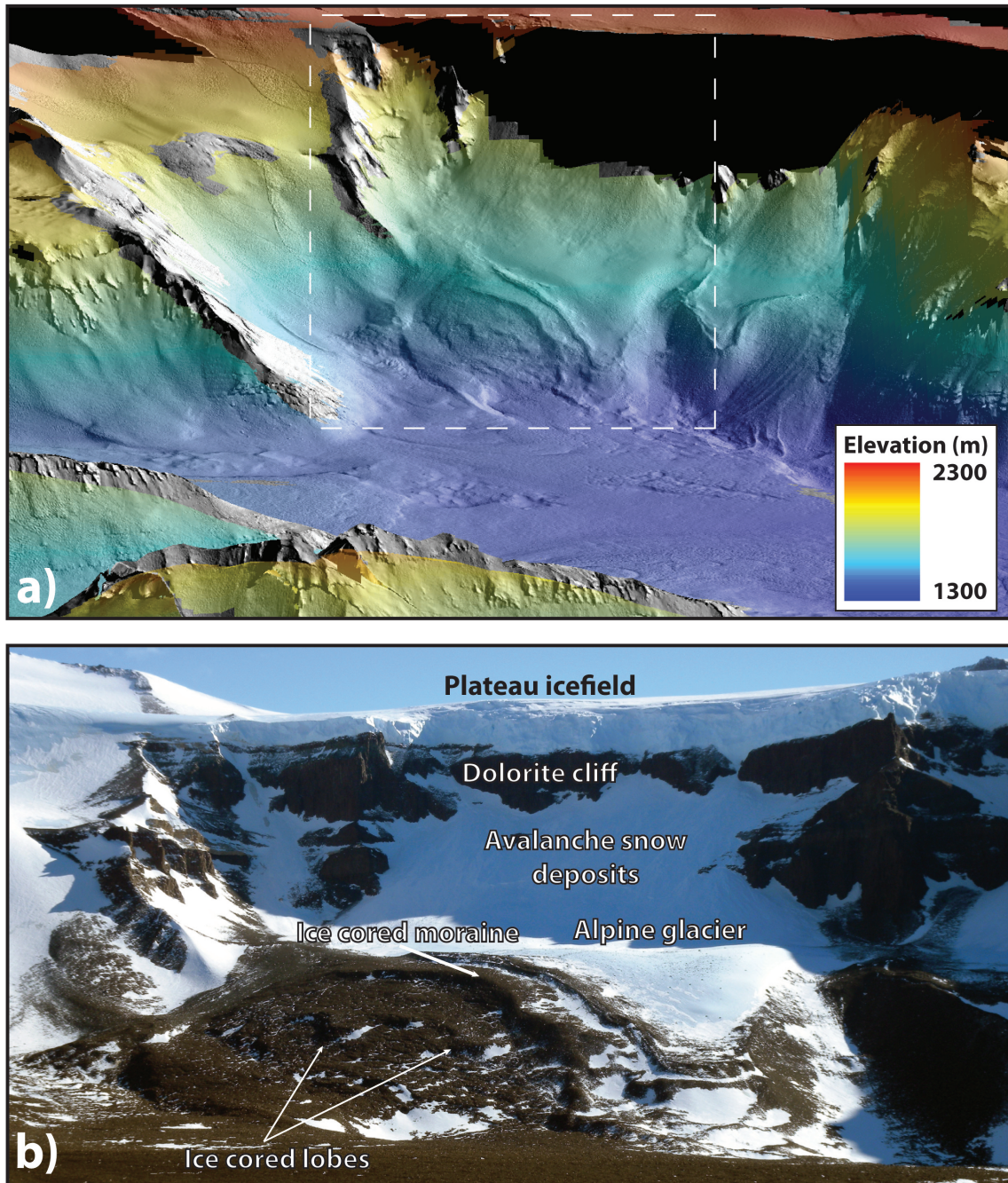
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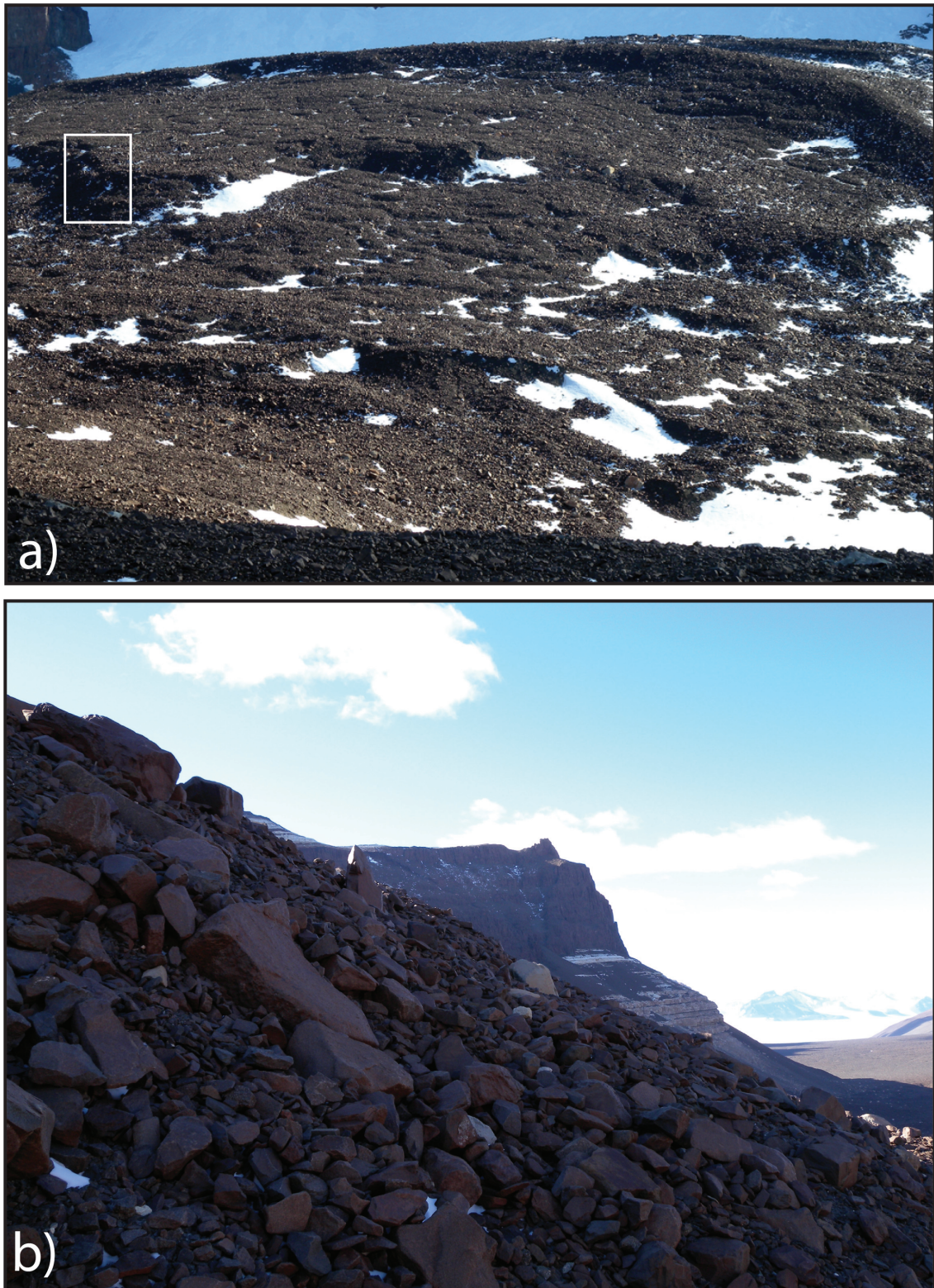
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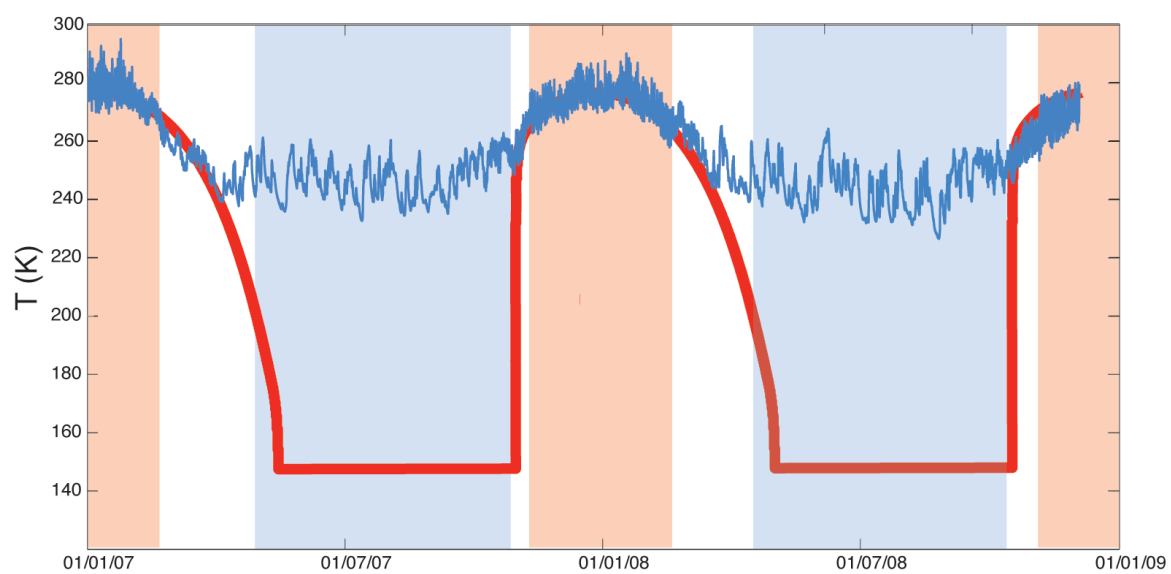
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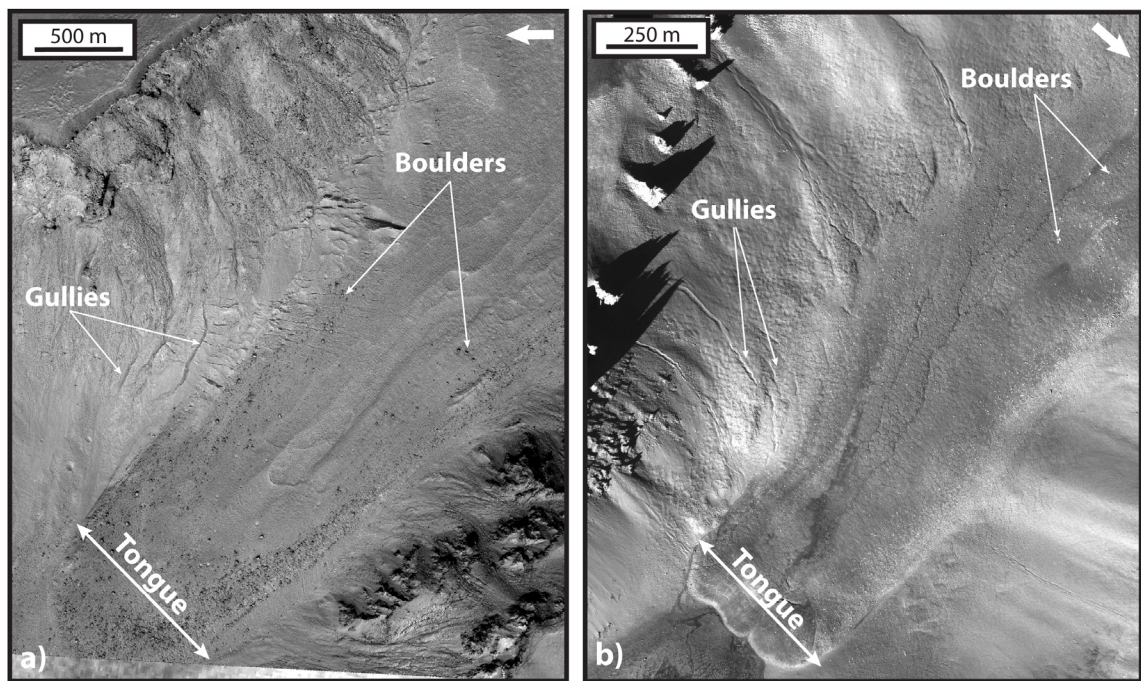
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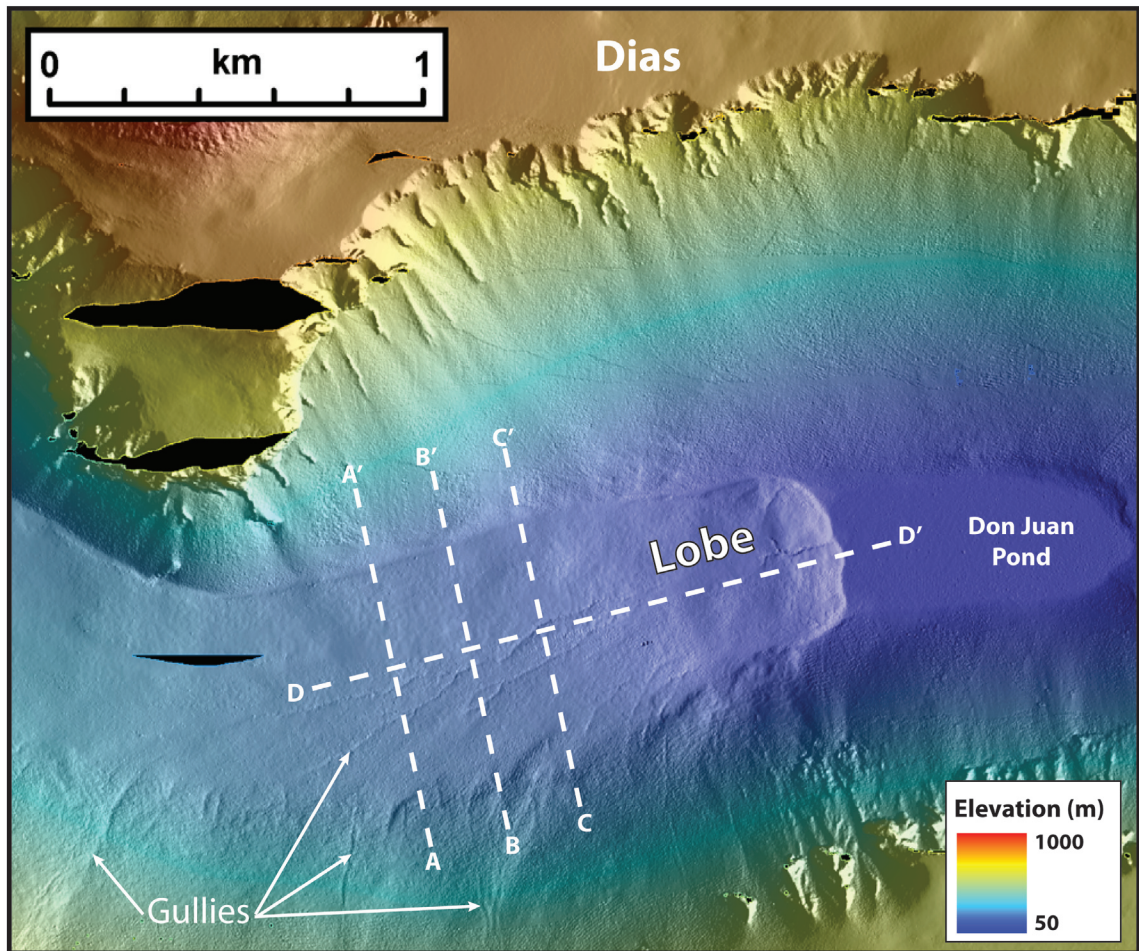
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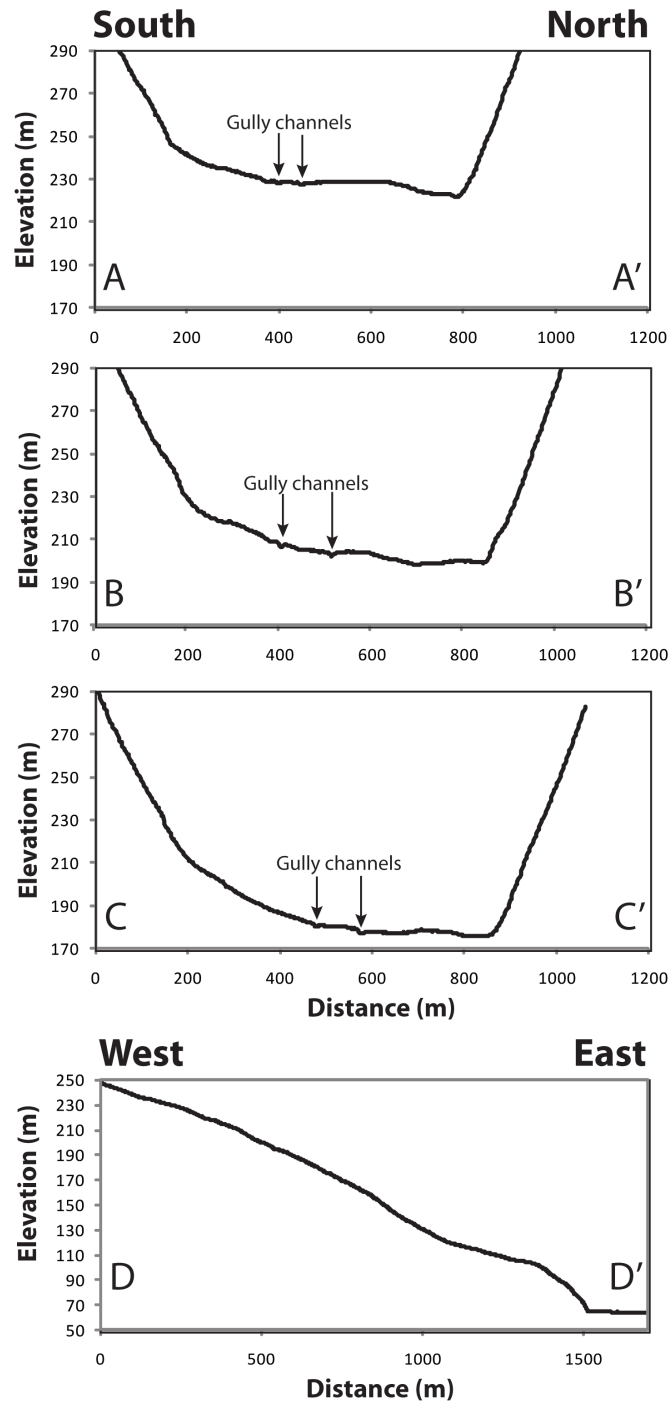
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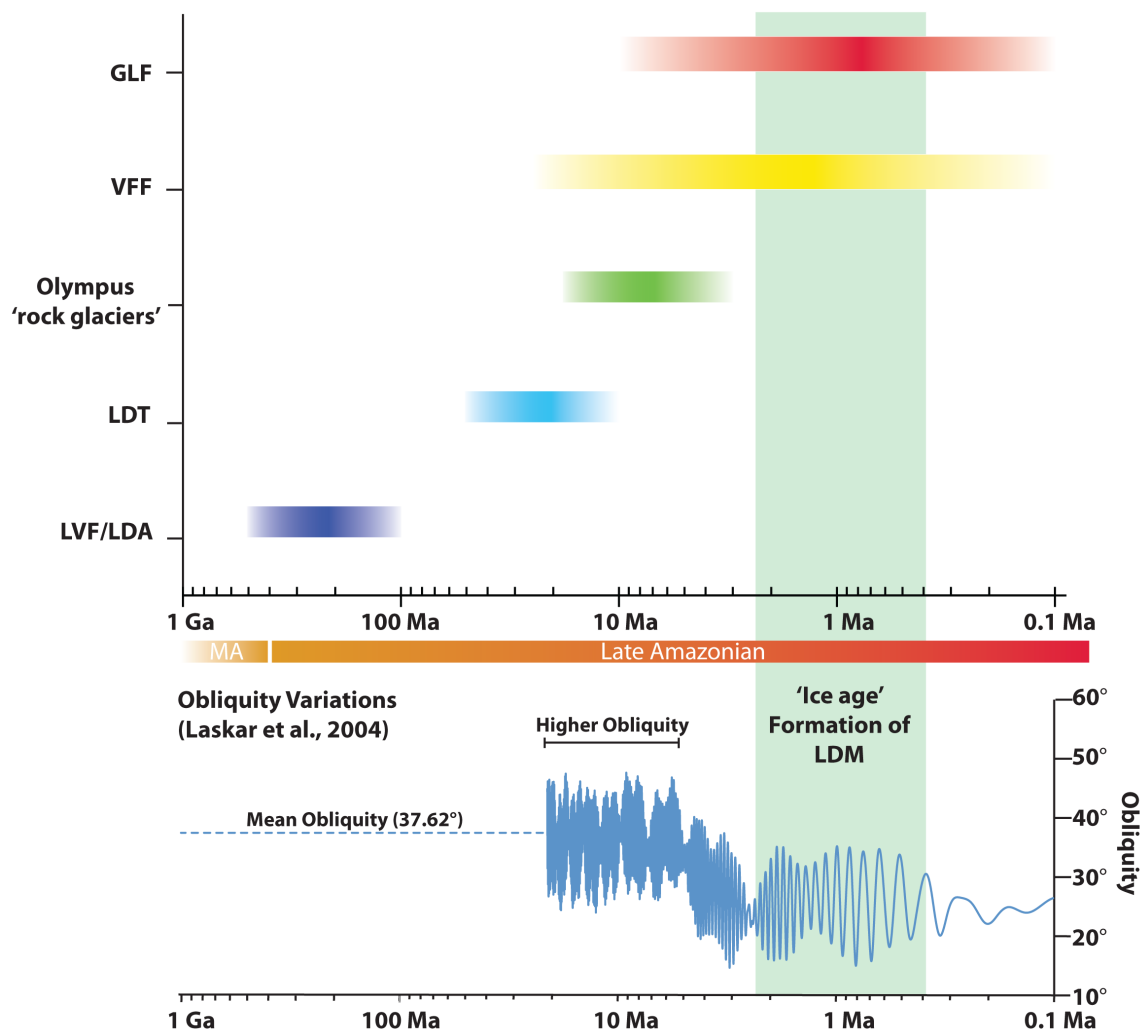
Chapter 6, Figure. 28



Chapter 6, Figure. 29



Chapter 6, Figure. 30



Chapter 6, Figure. 31

Chapter Seven

Synthesis: Post Noachian Mars Climate and Future Directions of Research

Gareth A. Morgan

1. Introduction

In the previous chapters of this thesis we have presented research into the geological history of two detailed study sites situated within each hemisphere of Mars. These studies have shown considerable evidence for climatic change on both a local and regional scale, demonstrating that post-Noachian Mars has experienced a complex climatic history. This chapter reviews the major conclusions drawn from the thesis research and raises outstanding questions that need to be addressed to further our understanding of the martian climate and its evolutionary history.

2. Hesperian Fluvial Activity

One of the most central lines of evidence to support the current paradigm that Mars underwent a transition from ‘warm and wet’ to ‘cold and dry’ conditions is the occurrence of large-scale valley networks suggestive of fluvial activity on Noachian terrains, and the relative rarity of these features elsewhere on Mars during later times. The morphological argument has been further supported by a new buffered crater-dating technique that places the most recent phase of large-scale fluvial activity just prior to the Noachian – Hesperian boundary (Fig. 1) (Fassett and Head, 2008). Mineralogical data derived from space-borne spectrometers have further emphasized this bi-climatic viewpoint of Mars history, showing clay-bearing species to be largely limited to Noachian terrain, and only anhydrous ferric alteration associated with Amazonian-aged regions (Bibring et al., 2006).

Mangold et al (2004) discovered Hesperian-aged highly dendritic valley networks close to Valles Marineris that, like the late Noachian valley networks, are suggestive of pluvial activity. The presence of these valleys was interpreted to represent either a gradual transition from the

warm and wet Noachian environment towards dry conditions through the Hesperian or alternatively that the change was episodic and that the Hesperian was punctuated with short, wet periods. In complete contrast to the Mangold et al. (2004) valleys, large-scale 10^3 kilometer-long outflow channels have also been identified as Hesperian in age. Many of these channels are interpreted to have been carved by the catastrophic release of groundwater due to pressure built up by an expanding cyrosphere (Carr, 1979). The formation of the outflow channels does not require warm and wet conditions; the possible involvement of a thick permafrost layer suggests that they formed while Mars was already very cold. Therefore, where do the Sinton valley networks presented in Chapter One fit between these two end-member classes of fluvial erosion, and what does this tell us about the Hesperian climate?

The morphology of the Sinton Valleys, consisting of anastomosing channels, flat floors, steep walls and low-order tributaries, in addition to the close spatial association with an impact crater, make them very different from the valleys identified by Mangold et al (2004). The Mangold et al. (2004) valleys are more consistent with terrestrial valleys arising from pluvial activity. The Sinton valleys also differ morphologically from outflow channels. The lack of theater-shaped alcoves or chaos-filled depressions at the source of the valleys and the occurrence of valleys close to the crest of topographic ridges suggests that the Sinton valleys were not sourced from groundwater. Hence, we propose that they originated from the thermal interaction between the ejecta from the Sinton crater-forming impact and a surface deposit of snow and ice. Morphologies associated with impacts into ice-rich substrates have long been suggested as the cause of fluidized ejecta on Mars (e.g. Carr et al., 1977). However, these differ from Sinton in that they have no associated valley networks. Therefore, what makes Sinton crater interesting from a climate standpoint is that the projectile that formed it appears to have impacted into surficial deposits of snow and ice, suggesting that there was sufficient ice cover on the plateau to form a plateau icefield and possible glacial activity during the Hesperian. This is consistent with other similar-sized Hesperian and Early Amazonian valleys located on volcanic edifices (e.g.

Ceraunius and Hecates Tholus). Fassett and Head (2006; 2007) argue that these formed from the melting of surficial deposits of snow and ice due to endogenic heating from intrusive magmatic sources in the edifices. We suggest that the presence of the Sinton valleys, in addition to the volcano valleys, provide evidence for the precipitation of snow and ice during the Hesperian. These valleys do not require a climate comprised of warm and wet conditions producing pluvial activity, but it does suggest that there was a redistribution of volatiles and deposition of surficial ice at latitudes where it is currently unstable. Heat sources, both endogenic (volcanic/magmatic) and exogenic (impact) are implied to have been sufficient to generate large volumes of melt to supply discharges $> 10^4 \text{ m}^3\text{s}^{-1}$ (Fassett and Head, 2007; Morgan and Head, 2009, Chapter One). This opens the possibility that smaller-scale melt features also formed from the available ice deposits. This may have included features similar to the Late Amazonian gullies (Malin and Edgett, 2000) that could have formed during periods of favorable insolation conditions. However, due to the degradation that would have occurred during the last three billion years, it would be unlikely that such subtle features would have been preserved until today.

3. Early – Middle Amazonian

The consistency between the morphological (Head et al., 2006a,b; Levy et al., 2007; Dickson et al., 2008; Morgan et al., 2009a) and radar (Holt et al., 2008 and Plaut et al., 2009) analysis of Lineated Valley Fill (LVF) and Lobate Debris Aprons (LDA) provides significant support for the interpretation that they are debris-covered glacial systems. This has important implications for our conception of the Amazonian climate as it suggests that massive volumes of ice were emplaced on the surface in the mid-latitudes in order to produce these glacial events.

In addition to the extensive LVF/LDA deposits along the northern portions of the Dichotomy Boundary, significant deposits of LDA also exist within the same latitudinal band (i.e. $30^\circ - 60^\circ$) in the southern Hemisphere (Squyres, 1979). Southern hemisphere LDA are not as

prevalent as in the north, with the exception of the Argyre and Hellas regions; this could be accounted for by the general lack of scarps and cliffs comparable to those associated with the fretted channels and the dichotomy boundary (Carr, 2006). The occurrence in both hemispheres argues for the global emplacement of ice at these latitude bands. Multiple studies have shown consistent age dates for the surface of LDA–LVF deposits in the range of ~100 – 500 Ma (e.g. Mangold, 2003; Levy et al., 2007, Morgan et al., 2009a). This implies that the last major phase of regional glaciation occurred within the Middle to Late Amazonian. Chapter Two provided evidence for more recent, smaller scale activity associated with small scale LVF lobes that superimpose, rather than feed, the larger scale integrated LVF–LDA systems. Similar observations have also been noted across the dichotomy boundary within Coloe Fossae (Dickson et al., 2008) and Nili Fossae (Levy et al., 2007), opening up the possibility that cyclic glaciation has been occurring throughout the Amazonian, but at a less significant scale than the LDA-LVF activity.

4. Late Amazonian – The last 100 Million Years

The most recent climatic transitions have been preserved in the greatest detail within the geomorphic record. The availability of high-resolution image data sets has permitted the investigation of the fine-scale stratigraphy present on the Martian surface that relates to the last several million years. Simulations of the orbital parameter variations of Mars over the last 20 Ma (Laskar et al., 2004) allow us to constrain the geomorphological interpretations within a timeframe of the insolation history. The recent developments in martian GCMs (e.g. Forget et al., 1999) can draw upon the Laskar et al (2004) simulations and provide a means to test geomorphic models (as was demonstrated in Chapter Five). In the following sections we first discuss the recent geological history in terms of the Laskar et al. (2004) obliquity simulations. This is followed by a discussion of the evidence for Late Amazonian change beyond 20 Ma.

It is the combination of obliquity, eccentricity and argument of perihelion that dictates the amount of insolation the surface of Mars receives in time and space. However, obliquity is the most dominant element (Carr, 2006) and thus the frequency of obliquity change has been viewed as a pacemaker for climatic change in many studies (e.g. Head et al., 2003; Milkovich and Head, 2004). The obliquity of Mars oscillates with a frequency of 1.2×10^5 yrs and the amplitude is modulated on a cycle of 2 Myr (Fig. 1). If one considers the longer-wavelength obliquity variation, we can see that the last 10 Ma can be divided into 4 broad periods (Fig. 1). Beginning in reverse chronologic order, the first period, and the one Mars is currently within, has lasted ~ 0.3 million years and is characterized by low obliquity values and stable polar ice caps that experience a seasonal expansion and recession over the high latitudes. This period could probably be described as the most ‘cold and dry’ and is not considered to have experienced the action of liquid water in any significant form, including the occurrence of ground thaw (Kreslavsky et al., 2008, Levy et al., 2009a). However, the identification of methane release from the surface of Mars in 2003 does argue for some level of geologic activity (Mumma et al., 2009). This could be due to current endogenic processes or interactions of ground water with basalt (serpentinization), or alternatively the result of recent processes that have facilitated the release of subsurface stores of methane (Mumma et al., 2009).

The second period corresponds to a collection of ~ 13 high-amplitude ($>30^\circ - <20^\circ$) fluctuations that occurred between $\sim 0.4 - 2.1$ Ma. This period has been termed an ‘ice age’ by Head et al. (2003) in which there is a net deposition of ice at the mid latitudes from the poles due to higher amounts of insolation at high latitudes driving the sublimation and migration of water from the water-ice polar caps. The redistribution of ice during this time is attributed to the formation of the ice-rich latitude-dependent mantle (Kreslavsky and Head, 2000; Mustard et al., 2001). The deposition of the mantle has also been associated with the formation of small viscous-flow features that have been identified across the martian surface at latitudes near 30° (Milliken et al., 2003). This is also the period in which pole-facing gullies are interpreted to have been last

active within the Asimov crater study site discussed in Chapter Five. This is consistent with other age estimates of gullies by crater counting techniques that pin the most recent activity to this period (Reiss et al., 2004; Schon et al., 2009).

Between 5 – 4 Ma the average obliquity value decreased from $\sim 35^\circ$ to 25° (Fig. 1). In addition to causing the surface/near surface stability of ice to change (Mellon and Jakosky, 1995), obliquity variations are also expected to have had an effect on the surface pressure, as greater volumes of CO₂ ice would be sublimated off the caps and into the atmosphere (Kieffer and Zent, 1992). The amount of pressure increase is a matter of debate (e.g. Mischna et al., 2003), but one would expect a notable climatic transition through this period in response to the decreasing obliquity, and it is something to be searched for in the geological record.

Prior to 4 Ma the obliquity of Mars was maintained at an average of $\sim 35^\circ$ degrees and the lowest values (25°) were above the present value. Maximum obliquity was in excess of 45° , and it is during these events that the equator-facing gullies within the Asimov crater study site (Chapter Five) are interpreted to have been active. The persistent high obliquity during this time would have had significant effects on the surface pressure and distribution of ice on Mars and may have been accompanied by relatively high levels of gully activity.

The remainder of the Late Amazonian is uncertain in terms of spin axis-orbital parameters, due to the non-deterministic nature of solar system dynamics prior to 20 Ma (Laskar et al., 2004). Nevertheless, as has been presented within this thesis (Chapters Two and Six) and the literature, ice-rich deposits did form during the last 100 Ma in the form of LDT (Chapter Six), glacial lobes on the flanks of Olympus Mons (Neukum et al., 2004; Head et al., 2005) and superimposed LVF lobes (Levy et al., 2007; Dickson et al., 2008; Morgan et al., 2009a). Statistical studies of obliquity variations over Mars history show that values as high as 82° are possible (Laskar et al., 2004). Therefore, through future investigations of the surface of Mars, it

may be possible to constrain obliquity variations beyond 20 Ma from the identification, location and dating of non-polar ice-rich deposits that are currently unstable on Mars (Head et al., 2009).

5. Future Questions.

This section outlines some of the outstanding questions that have arisen from the research presented in this thesis and briefly reports on some of the initial work undertaken to address these issues.

5.1 Impact-Induced Valley Formation

The Sinton valley model represents a special case of the impact-induced valley formation model first proposed by Brackenridge et al. (1985). Brackenridge et al. (1985) suggested that hydrothermal activity initiated by impact events could account for valley formation through sapping and thus negate the need for a warmer Noachian climate. From our analysis presented in chapter one, we disagree that this argument can be applied to the Sinton model, because the resulting meltwater production associated with the impact is interpreted to have been too rapid and focused to permit the formation of valleys similar to those present within Noachian terrain. However, if ice was periodically deposited along the dichotomy boundary during the Hesperian and into Amazonian (as the presence of LVF/LDA deposits discussed in Chapter Two suggests), it is possible that other impacts also interacted with surficial deposits of snow and ice. Therefore, detailed surveys should be carried out of all craters > 10 km diameter within the 30° - 50° latitude band in order to search for the presence of similar valley networks similar to those seen at Sinton. This would establish whether the valleys associated with Sinton crater represent a unique fluvial assemblage or whether impacts provided a source of meltwater throughout the Hesperian – Amazonian epoch.

5.2 Lineated Valley Fill and Lobate Debris Aprons

In order for the full extent of Amazonian glacial history to be realized we need to address whether the topography of the dichotomy boundary was essential for the formation of LVF/LDA, or whether the availability of debris from the steep, >1 km-high fretted valley cliffs was essential only in the preservation of the ice deposits. Due to the distinct lack of erosion associated with cold-based glaciation, it is possible that much larger bodies of ice existed in both hemispheres, but that their presence is difficult to detect due to the lack of distinctive remaining deposits. Initial studies of the northern plains surrounding the termini of LDA and LVF deposits with MOC revealed the presence of unusual 1 km wide, layered deposits (Carr, 2001). We followed up Carr (2001) observations with CTX and HiRISE and found these deposits to be typically draped along the edge of small cliffs or crater walls, with the layers orientated at the angle of dip (Fig. 2) (Morgan and Head, 2008). The origin of the deposits is uncertain, and due to their size it is not possible to provide an age from the size frequency distribution of superimposed craters. However, their spatial association with LVF/LDA deposits suggests that the two features are related. One possibility is that they represent the remnants of glacial material (debris or ice-cored debris) deposited during the retreat of large ice sheets that extended into the northern plains. Addressing this hypothesis is of significant importance to Amazonian climate research and thus more attention should be devoted to the study of these features in the future.

5.3 Gullies and the Late Amazonian.

Chapters Three to Five outlined the major work that has been undertaken by a range of authors in the investigation of gullies since their discovery on Mars by Malin and Edgett (2000). Despite the progress made, three major questions still remain to be fully answered:

1. *Can the melting of snow account for the formation of all gullies on Mars?* Chapters Three to Five discussed the development and application of a snowmelt model of gully formation

from fieldwork conducted in Antarctica. This model assumed that snow was both generated during periods of high obliquity on Mars and that it could be redistributed and concentrated in topographic hollows to form snowbanks. This assumption has also been used in other published work on gullies (Williams et al., 2009) and so needs to be tested to determine whether snow could provide a source of water at every gully site on Mars. One method is through the application of GCMs that could model the activity and behavior of wind over DTMs of the martian surface. In addition to this, wind tunnel experiments involving snow deposits under martian conditions could also be applied.

2. What is the ratio of water to debris during gully formation? The gully activity observed in the Dry Valleys was largely in the form of fluvial erosion involving surface runoff, but several authors have suggested that some gully activity results from the initiation of debris flows with water content as low as $\sim 10\%$ (e.g. Costard et al., 2002). Recent research that I have been involved with has identified small-scale lobate flow-fronts along the edge of gully fans at a study site in the northern hemisphere (Levy et al., 2009b). These features may represent the viscous deposits of debris flows. If this interpretation is correct these features represent an end-member type of gully that is very different than the fluvial erosional forms seen in gullies elsewhere on Mars (McEwen et al., 2007). This opens up the possibility that different types of erosion, involving varying amounts of water, have been responsible for gully formation on Mars.

3. What is the timing and duration of gully activity on Mars? Chapter Five reported on the results of models of potential gully activity that suggested gullies on slopes of different orientations had formed over different time periods. This is supported by other studies that have dated the most recent gully activity to ~ 1 Ma (Reiss et al., 2004; Schon et al., 2009). However, much uncertainty still remains over how long gullies have been forming prior to this date. Have they been active throughout the Amazonian or are they a recent phenomenon? Insight into the potential longevity and variation of gully activity was provided during a revisit to the Dry Valleys gully site discussed in Chapter Three and Four during the austral summer of 2008-9. Extremely

low amounts of precipitation had starved the gullies of snowbanks, preventing the occurrence of the type of meltwater generation and channel erosion observed in 2006 (Fig. 3) (Morgan et al., 2009b). This suggests that martian gullies may also be extremely sensitive to variations in their local microclimatic conditions and that the level of fluvial activity may have undergone large variations with time.

6. Conclusion.

The research presented in this thesis is consistent with the present climatic paradigm, which is that Mars has not experienced widespread pluvial activity since the end of the Noachian. Nevertheless, the expression ‘cold and dry’ is only useful in a relative sense and does not adequately define the level of complexity the Martian climate has experienced as a result of large amplitude variations in obliquity and eccentricity. Evidence for surface deposits of glacial ice is widespread in both hemispheres at latitudes lower than the current extent of the seasonal polar caps. Therefore, calculations of the volume of the global reservoir of water on Mars that only include the polar caps should be considered an underestimate.

In contrast to the complexity of the Earth’s climate system, the absence of oceans and the thin atmosphere present on Amazonian Mars highlights the planet as a natural laboratory for climate researchers to test the fundamental global flux of water in its three phases in response to Milankovitch cycle forcing. The presence of non-polar ice that has been preserved over 10^2 Ma (Head et al., 2009), will provide a unique record of global change for future missions to investigate and provides a potential resource of water for future human exploration of Mars.

In the absence of rainfall, the generation of meltwater from snow and ice provides an alternative source of water that can be utilized by geomorphic processes operating on the martian surface. Under such environmental conditions, water is limited to specific locations that are either areas of preferential insolation conditions (microclimates, i.e. gully formation) or they are characterized by an alternative heat source from endogenic (Fassett and Head, 2006; 2007) or

exogenic processes (Brakenridge et al., 1985; Morgan and Head, 2009). Nevertheless, life in water-limited terrestrial environments such as the Antarctic Dry Valleys have developed strategies for coping with long periods of desiccation and can respond rapidly to inter-annual sources of meltwater (e.g. McKnight et al., 1999). Whether this can or has happened on Mars is of course an open question, but we do know that water, the essential ingredient for life has been present on the surface over recent history.

The NASA Mars Program mission has been to ‘follow the water’, and I believe it has been successful in that objective. Justified concerns over the cross-contamination of Earth and Mars do limit our ability to explore the most recent sites of liquid water. However, as the Phoenix Lander demonstrated in 2008, there is also scientific value in ‘following the ice’. The large non-polar ice deposits investigated in Noachis Terra (Chapter Six) and the northern Dichotomy boundary (Chapter Two) in this thesis account for only a few of the locations that ice may still exist close to the surface today. Therefore, I propose that engineering new missions that can access and sample this ice is a valid direction for future missions.

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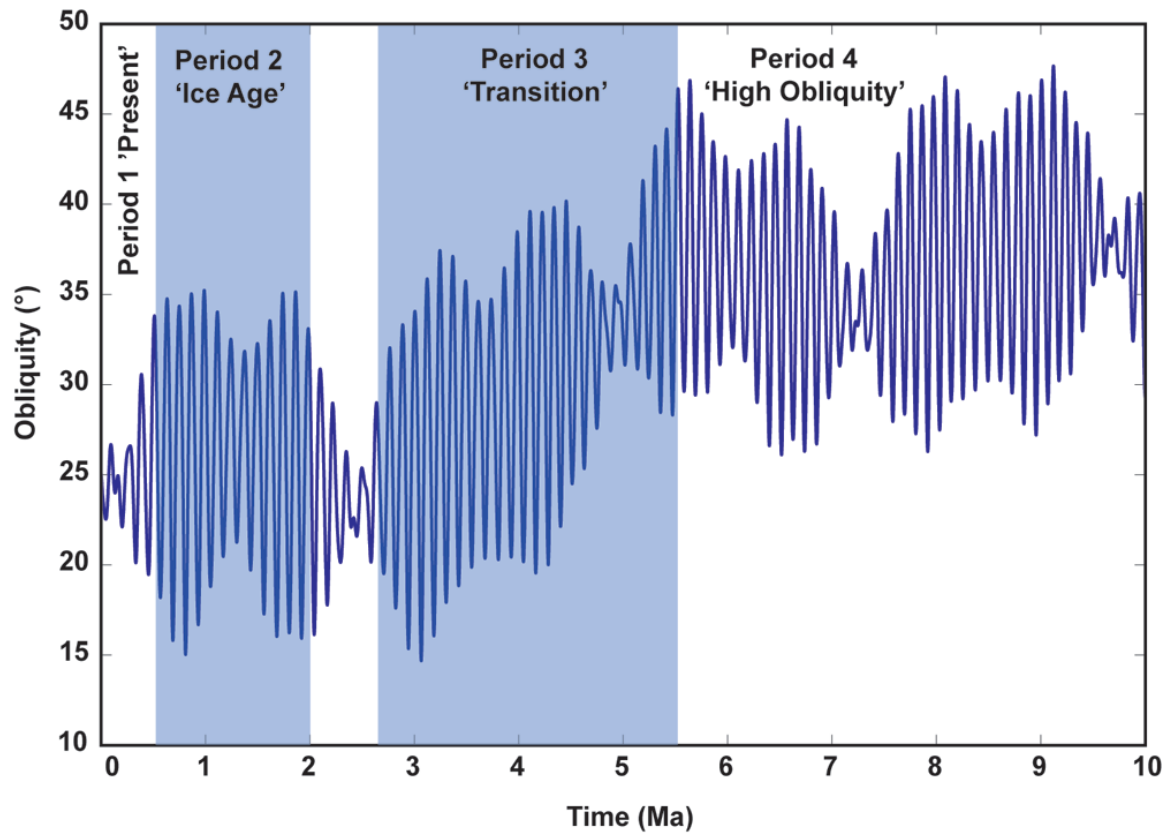
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Figure Captions

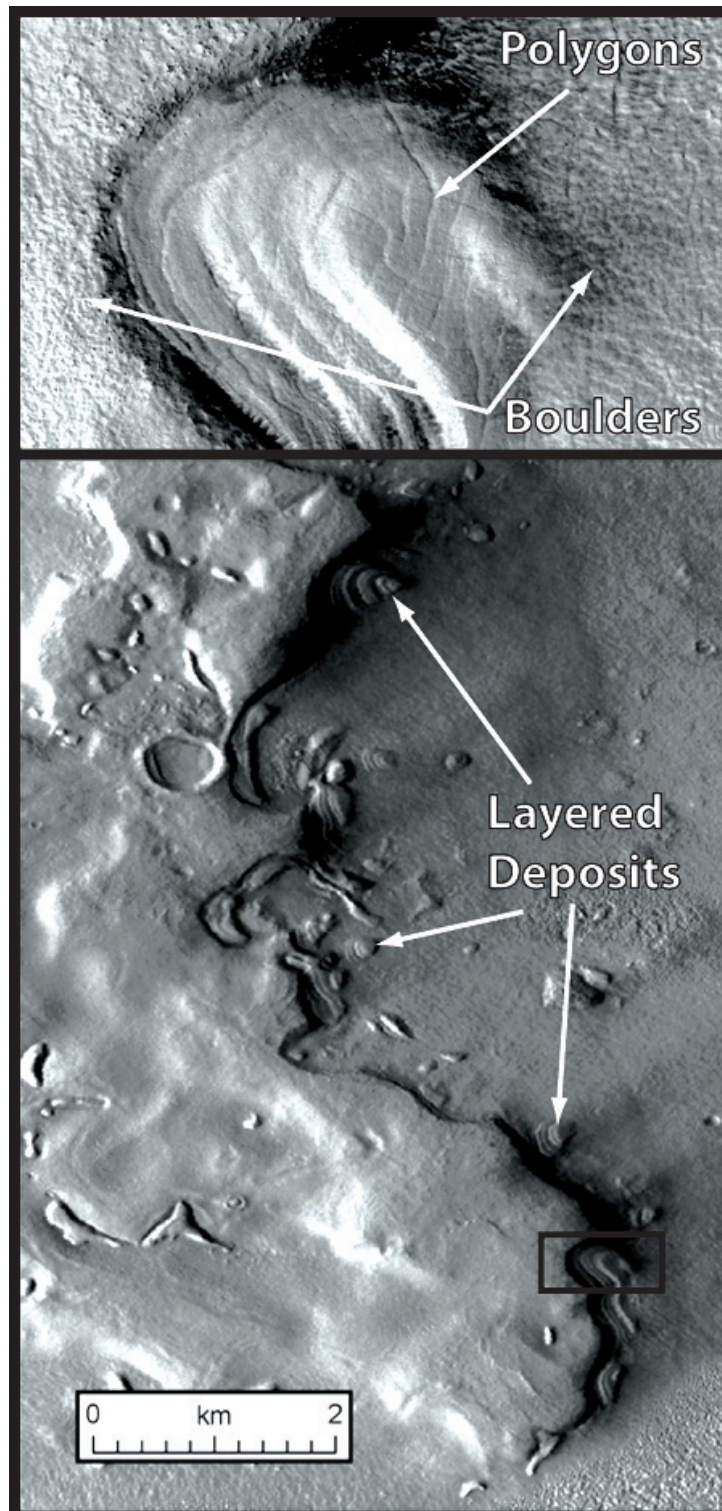
Fig. 1. Laskar et al (2004) plot of martian obliquity variations over the last 10 Ma. The long-wavelength variations in the amplitude of obliquity are highlighted and are discussed in the text in terms of their implications for the Late Amazonian climate.

Fig. 2. Layered deposits in the northern plains adjacent to regions of the northern dichotomy boundary in which LVF and LDA are present. The deposits may have been emplaced during the retreat of former ice sheets extending from the Dichotomy Boundary. Lower image CTX: P02_001650_2252. Inset: HiRISE: PSP_006674_2230.

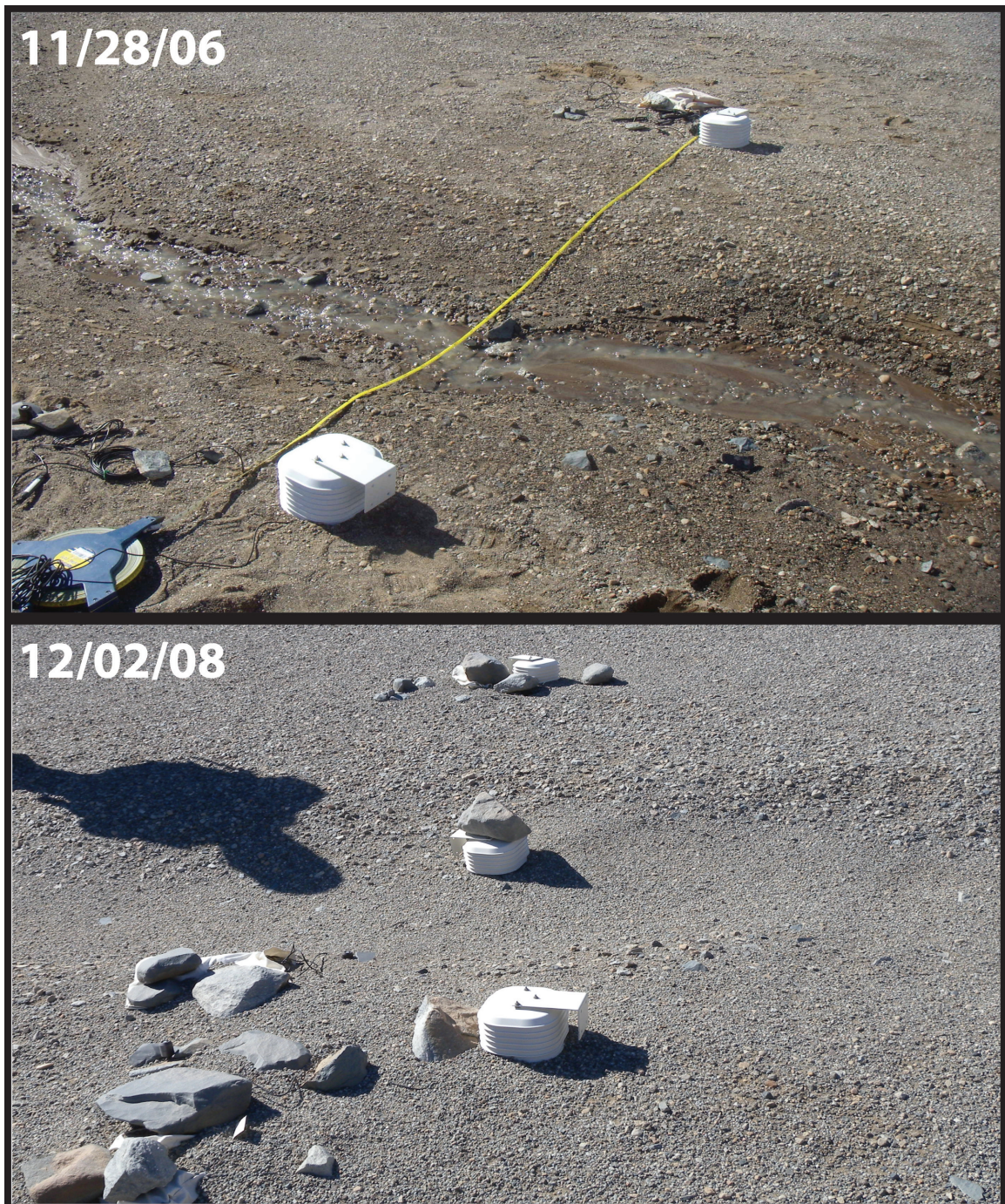
Fig. 3. Difference in fluvial activity within a gully system in the Antarctic Dry Valleys between the 2006 (upper image) and 2008 (lower image) field seasons as a result of the lack of snow in 2008 relative to 2006.



Chapter 7, Figure. 1



Chapter 7, Figure. 2



Chapter 7, Figure. 3

Instrument Name/Data Sat	Mission	Data Set Type	Resolution	Coverage of Mars	Specific use
Imager	Viking	Visible Imagery	150-300m/pixel	Global	Context imagery, global surveys, geological mapping, crater counting.
Mars Orbital Camera (MOC)	Mars Global Surveyor	Visible Imagery	1.5m/pixel	~ 5%	Very high-resolution studies of the surface.
Mars Orbital Laser Altimeter (MOLA)	Mars Global Surveyor	Digital Elevation Models	Foot print: 460m. Vertical: 30 cm	Global	Utilized to produce Digital Elevation Models (DEM), from which topographic profiles and slope maps can be produced.
Thermal Emission Imaging System (THEMIS)	Mars Odyssey	Visible Images	19m/pixel	>50%	High-resolution studies of the surface.
Thermal Emission Imaging System (THEMIS)	Mars Odyssey	Infrared Images	100m/pixel	Global	Nighttime Images: Quantify thermal inertia of surface materials. Daytime Images: Context Images, global surveys, geological mapping and crater counting.
High Resolution Stereo Camera (HRSC)	Mars Express	Visible color and stereo Images	10m/pixel	>50%	High resolution studies of the surface, geological mapping, crater counting.
High Resolution Stereo Camera (HRSC)	Mars Express	Digital Elevation Models	100 - 200m/pixel	Restricted to area covered by images	Utilized to produce Digital Elevation Models (DEM), from which topographic profiles and slope maps can be produced.
High Resolution Imaging Science Experiment (HiRISE)	Mars Reconnaissance Orbiter	Visible Images	30cm/pixel	>1%	Highest resolution studies of the surface.
Context Camera (CTX)	Mars Reconnaissance Orbiter	Visible Images	6m/pixel	>35%	High resolution studies of the surface.

APPENDIX 1. Mars Data Sets