

Evaluating the Hydrology of a Cold and Icy Early Mars

By

James P. Cassanelli

B.S. Geology and Geophysics, University of Connecticut, 2009

M.Sc. Geology and Geophysics, University of Connecticut, 2011

M.Sc. Geological Sciences, Brown University, 2015

A dissertation submitted in partial fulfillment of the requirements for the degree of
Doctor of Philosophy in the program of Earth, Environmental, and Planetary Sciences at
Brown University

Providence, Rhode Island

May, 2019

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This dissertation by James P. Cassanelli is accepted in its present form by the Department
of Earth, Environmental, and Planetary Sciences as satisfying the dissertation
requirement for the degree of Doctor of Philosophy

Date _____

James W. Head, Brown University, Advisor

Recommended to the Graduate Council

Date _____

Edgar M. Parmentier, Brown University, Reader

Date _____

James M. Russell, Brown University, Reader

Date _____

Malcolm J. Rutherford, Brown University, Reader

Date _____

James L. Fastook, University of Maine, Reader

Approved by the Graduate Council

Date _____

Andrew G. Campbell, Dean of the Graduate School

James P. Cassanelli – Email: James.Cassanelli@gmail.com

Home Page: http://www.planetary.brown.edu/grad_pages/Cassanelli/

LinkedIn: <https://www.linkedin.com/in/james-cassanelli-ab02b4127>

Google Scholar (publication list): <https://scholar.google.com/citations?user=wwfKWxsAAAAJ&hl=en>

EDUCATION & RESEARCH EXPERIENCE

Brown University: Ph.D. in Geological Sciences

2013 – Present

Brown University: M.Sc. in Geological Sciences

2013 – 2015

Emphasis on: Planetary science, geophysics

Thesis topic: Evaluation of the hydrology of early Mars.

- Geophysical and numerical analysis of planetary geologic processes including heat flow, fluid flow, porous media flow, and fluvial processes.

University of Connecticut: M.Sc. in Geology and Geophysics

2009 – 2011

University of Connecticut: B.S. in Geology and Geophysics

2005 – 2009

Emphasis on: Hydrogeology, geology, geophysics

Thesis topic: Assessment of spatial and temporal trends in the salinity of Connecticut's groundwater.

- Regional scale GIS analysis of groundwater salinity supported by field scale sampling and monitoring.
-

PROFESSIONAL EXPERIENCE

Geosciences Intern – Anadarko Petroleum Corporation

May 2018 – August 2018

- De-risked a new ventures Cretaceous basin floor fan play through a comprehensive, basin-wide analogue analysis.
- Developed an integrated basin depositional model through regional interpretation of 2D seismic and well data.
- Collaborated with subject matter experts in petrophysics, geochemistry, structural geology, and sedimentology/stratigraphy to refine and test source and reservoir provenance models.
- Provided commercial overview of analogue basin and outlined recommendations for business decisions in company acreage.

Hydrogeologist – Leggette, Brashears, and Graham Inc.

June 2010 – April 2013

- Collected and documented critical data for water supply exploration and environmental remediation projects.
 - Supervised drilling and testing of water supply, monitoring, and remediation wells.
 - Conducted data and sample analysis and prepared technical reports for clientele and regulatory agencies.
 - Worked in team-based environment, assisted in training of junior staff.
-

COMPUTATIONAL EXPERIENCE & RELEVANT SKILLS

Languages: MATLAB, Python, SQL, HTML, CSS (some exp.: C++, Fortran)

Programs: ArcGIS, MS Office suite (including Access), ADOBE creative suite, Mathematica, ERDAS Imagine, ENVI, DecisionSpace Geosciences

- Experienced in scientific computing, numerical modeling, construction and application of geologic models.
- Skilled in utilization, visualization, and interpretation of geologic, geophysical, and remote sensing data sets (1-D/2-D/3-D).
- Significant experience in technical writing/communication including publication of several papers and presentations at >10 national/international science conferences.
- Practical and professional experience in: geochemical/geophysical well profiling, bore logging, subsurface lithology and poroperm characterization, soil/water sample collection, tracer studies, well planning and design.

FIELD EXPERIENCE

Sedimentary field trip: Permian Basin, New Mexico and West Texas	<i>June 2016</i>
3 rd Nordic-Hawaii NASA Astrobiology Institute Summer School: Iceland.	<i>July 2015</i>
Brown University geology summer field trip: Nova Scotia, Canada.	<i>August 2013</i>
Hydrogeology consultant field experience.	<i>2010 – 2013</i>
Field course in structural geol., sedimentol., and geomicrobio.: Puerto Rico.	<i>March 2008</i>
Field problems in Earth structure: New York and Vermont.	<i>October 2007</i>

ACADEMIC AWARDS & CERTIFICATIONS

Lunar Planetary Institute Career Development Award.	<i>March 2018</i>
Rocky Mtn. Rendezvous (UWYO) – 2 nd Place Poster Award.	<i>September 2017</i>
Brown University Dissertation Fellowship recipient.	<i>July 2017</i>
Sigma Xi Honor Society Nominee.	<i>February 2016</i>
Brown University First Year Fellowship recipient.	<i>June 2013</i>
EPOC scholarship recipient in environmental sciences.	<i>September 2009</i>
University of Connecticut Dean's List for academic performance.	<i>2008, 2009</i>
New England Scholar for academic excellence.	<i>2008</i>

PROFESSIONAL AFFILIATIONS AND SERVICE

Reviewer for: JEQ.	<i>2017 – Present</i>
Elsevier Earth and Planetary Science book proposal reviewer.	<i>March 2017</i>
American Geophysical Union (AGU) – Student Member.	<i>2017 – Present</i>
American Association of Petroleum Geo. (AAPG) – Student Member.	<i>2017 – Present</i>
Society of Exploration Geophysicists (SEG) – Student Member.	<i>2017 – Present</i>
Society for Sedimentary Geology (SEPM) – Student Member.	<i>2017 – Present</i>
Brown University Geology Mentor Representative.	<i>2016 – 2018</i>
Brown University Geology Club President.	<i>2015 – 2016</i>
Geological Society of America (GSA) – Student Member.	<i>2007 – 2009</i>

PUBLICATION LIST

Peer-reviewed Publications

(Google Scholar profile: <https://scholar.google.com/citations?user=wwfKWXsAAAAJ&hl=en>)

- Cassanelli, J.P., and Head, J.W., (2018). Large-scale lava-ice interactions on Mars: Investigating its role during Late Amazonian Central Elysium Planitia volcanism and the formation of Athabasca Valles. *Planetary and Space Science*, 158, 96-109.
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- Cassanelli, J.P., and Head, J.W., (2016). Did the Orientale impact melt sheet undergo large-scale igneous differentiation by crystal settling? *Geophysical Research Letters*, 43(21), 11156-11165.
- Cassanelli, J.P., and Head, J.W., (2016). Lava heating and loading of ice sheets on early Mars: Predictions for meltwater generation, groundwater recharge, and resulting landforms. *Icarus*, 271, 237-264.
- Cassanelli, J.P., and Head, J.W., (2015). Firm densification in a late Noachian “icy highlands” Mars: Implications for ice sheet evolution and thermal response. *Icarus*, 253, 243-255.
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- Cassanelli, J.P., and Robbins, G.A., (2013). Effects of road salt on Connecticut’s groundwater: A statewide centennial perspective. *Journal of Environmental Quality*, 42(3), 737-748.

Selected Conference Proceedings

- Cassanelli, J.P., and Head, J.W., (2018). Assessing the formation of valley networks on a cold early Mars: Predictions for erosion rates and channel morphology. *49th Lunar and Planetary Science Conference*, Abstract 1124.
- Cassanelli, J.P., and Head, J.W., (2018). Evaluating the role of large-scale lava-ice interactions during Elysium Planitia volcanism and the formation of Athabasca Valles. *49th Lunar and Planetary Science Conference*, Abstract 1125.

- Cassanelli, J.P., and Head, J.W., (2018). Impact melt sheet composition, age, and igneous differentiation?: Commercial mission goals. *Lunar Science for Landed Missions Workshop*. Abstract LLW2018-28.
- Cassanelli, J.P. (2017). Formation of valley networks in a cold and icy early Mars climate: Predictions for erosion rates and channel morphology. *2017 American Geophysical Union Fall Meeting*.
- Cassanelli, J.P., and Head, J.W., (2017). Formation of valley networks in a cold and icy early Mars climate: Predictions for erosion rates and channel morphology. *4th Conference on Early Mars*. Abstract 3020.
- Cassanelli, J.P., and Head, J.W., (2017). Formation of the Reull Vallis outflow channel by large-scale lava-ice interactions and top-down melting. *48th Lunar and Planetary Science Conference*, Abstract 1188.
- Cassanelli, J.P., and Head, J.W., (2017). Valley network formation in a “cold and icy” climate regime: Theoretical predictions for erosion rates and channel morphology. *48th Lunar and Planetary Science Conference*, Abstract 1191.
- Cassanelli, J.P., and Head, J.W., (2016). Did the Orientale impact melt sheet undergo large-scale igneous differentiation by crystal settling? *7th Planetary Cratering Consortium*.
- Cassanelli, J.P., and Head, J.W., (2016). Did the Orientale impact melt sheet undergo large-scale igneous differentiation by crystal settling? *47th Lunar and Planetary Science Conference*, Abstract 1174.
- Cassanelli, J.P., and Head, J.W., (2016). Lava heating and loading of ice sheets on early Mars: Predictions for meltwater generation, groundwater recharge, and resulting landforms. *47th Lunar and Planetary Science Conference*, Abstract 1176.
- Cassanelli, J.P. et al., (2015). Ice sheet basal melting on Mars: Potential for long-term habitats. *3rd Nordic-Hawaii NASA Astrobiology Institute Summer School, Iceland*.
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- Cassanelli, J.P., and Head, J.W., (2015). Lava-loading of ice sheets in a late Noachian “icy highlands” Mars: Predictions for meltwater generation, groundwater recharge, and resulting landforms. *46th Lunar and Planetary Science Conference*, Abstract 1533.
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- Cassanelli, J.P., and Head, J.W., (2014). Firn densification on late Noachian Mars: Implications for ice sheet formation and thermal characteristics. *8th International Conference on Mars*, Abstract 1032.
- Cassanelli, J.P., and Head, J.W., (2014). Firn densification on late Noachian Mars: Implications for ice sheet formation and thermal characteristics. *45th Lunar and Planetary Science Conference*, Abstract 1265.
- Cassanelli, J.P., and Robbins, G.A., (2010) Evaluating impacts of anthropogenic salt influxes to Connecticut’s groundwater through water quality monitoring and tracer studies. *45th Geological Society of America Northeastern Section Meeting*, Abstract 57-11.

Acknowledgements

Throughout my time at Brown, I often reminded myself of the phrase “nothing easy is truly worth doing.” While my academic endeavors have been anything but easy, they were unquestionably made *easier* by a great many people whose invaluable help I would like to acknowledge.

First, I must acknowledge the remarkable efforts of my advisor Jim Head. Jim’s unparalleled enthusiasm for planetary science, deep wisdom and experience, steadfast support and encouragement, thoughtful guidance, and impeccable beverage selection enabled every facet of my scientific development and success in my graduate pursuits. My deepest gratitude is owed to Jim for the opportunities that he has afforded to me, it has been a sincere pleasure to have worked with Jim as an advisor and a friend.

Many thanks are due to the members of my preliminary and dissertation advisory committees. The time spent in meetings and discussions as well as the insightful guidance and suggestions provided by Marc Parmentier, Jim Russell, Mac Rutherford, Jung-Eun Lee, and James Fastook have all been greatly appreciated.

To the many faculty members who contributed to my education and development as a research scientist, my sincere thanks for the effort, time, and enthusiasm brought forth in each and every class (and field trip). Special thanks to Ralph Milliken, Marc Parmentier, Jim Russell, Don Forsyth, Greg Hirth, Steve Parman, Jung-Eun Lee, Brandon Johnson, and Chris Huber for exceptionally instructive courses and discussions.

Thanks also to the members of the administrative and technical staff: Melissa DeAugustinis, Anne Côté, Lisa Noble, Pat Davey, Bill Collins, and Dave Blair. The logistical work put into facilitating conference travels and activities as well as into

ensuring the operation of various departmental functions and equipment is gratefully acknowledged. Particular thanks are extended to Jay Dickson whose instrumental technical support and ArcGIS savvy allowed the efficient operation of the entire sea-floor/mission room research team.

The many friends I have made within and outside the DEEPs department have truly enriched my experience at Brown, especially my fellow sea-floor lab, mission room, and Charlestown Chiefs members. Thanks to the senior graduate students, Jennifer Whitten, David Baker, Tim Goudge, Lauren Jozwiak, and Kat Scanlon who provided helpful advice when I was a mere youngling. To my contemporary Erica Jawin: from start to finish, we went through it all at together, beige food, austere Russian McDonalds, prelim stress, and victory dances. You were a most excellent partner in crime (*finger guns*). To Ariel Deutsch: I don't know if anyone will ever understand our mutual struggles roughing it through the 77A life, but I'm certainly glad to have had such a friend to have made it through with. To Ashley Palumbo: thanks for all the laughs through your lab sass and persistent bullying, and also for the delicious radlers at your wedding. To Kevin Cannon: thanks for always being a willing, and often instigating, participant in poor beverage selections, for your patience with my Icelandic transgressions, and for putting together some seriously excellent ski trips. To Chris Kremer: you might be new on the scene, but I can already tell you have a bright future in planetary science ahead of you. Thanks for all the career advice and pro tips, they have proven to be exceptionally handy. To David Weiss: last but certainly not least, thanks David for your great friendship. Our mutually beneficial, near-symbiotic research efforts were academically invigorating and always a source of fascinating, enlightening discussion. Equally enjoyable was our time

spent as roommates, where your comparably stringent dietary and exercise regimes helped inspire me to stick with mine, and the evening anime/gaming sessions and general roommate antics were always a source of needed entertainment.

Finally, I would like to acknowledge and thank my family without whose constant encouragement and support this all would not have been possible. Mom and Dad, I could never thank you enough for everything you have done for me, for always being there to lend an open ear or helping hand through difficult times and a loving presence to share the good ones.

Preface

A primary motivation for the exploration of the planetary bodies within our solar system is to determine if, where, and when life may have existed beyond Earth, and the nature of the habitats that supported it. Given the critical role of liquid water in the sustenance of life as it is currently understood, the presence of liquid water has served as a practical minimum criterion to establish the existence of a habitable environment. Therefore, a fundamentally important aspect of planetary exploration efforts has been to identify and characterize environments which hosted liquid water. This is particularly true of the exploration of Mars, which for more than two decades has been carried out through a robust NASA program with a prime directive to “follow the water.”

Early exploration of Mars revealed encouraging indications of a potentially Earth-like active hydrologic system early in the planets history during the Noachian period. However, despite decades of subsequent focused study, many questions regarding the nature of the hydrologic system of early Mars remain unresolved. Of particular importance is the fundamental question whether the early Mars hydrologic system was characterized by cold and icy, or warm and wet conditions, the subject on which this dissertation concentrates. The main objective of this dissertation is to determine if a characteristically cold and icy early Mars hydrologic system, the condition predicted by recent climate modeling studies, is consistent with the presence and morphology of the wide array of hydrologic features preserved on the planet’s surface. While Mars is host to a wide range of hydrologic features, the examination presented in this dissertation focuses primarily on the fluvial features. The fluvial features of Mars include the valley

networks and outflow channels, arguably the most spectacular water-formed features on the planet.

This dissertation builds upon the work of two separate past investigations which constituted a masters thesis. The first of these investigations evaluated in detail the thermophysical properties of regional ice sheets predicted to form on Mars as a result of cold icy conditions climates conditions. This assessment was performed in order to constrain the response of martian ice sheets to top-down and bottom-up heating processes. Results of this analysis indicated that under a wide range of plausible ice accumulation and climate conditions, basal melting of martian ice sheets required anomalously large geothermal heat fluxes, highly asymmetrical ice accumulation producing exceptionally large local ice thicknesses, or a coeval component of sustained top-down heating. The second investigation explored the implications of the accumulation of martian ice sheets within volcanically active terrains, specifically the Tharsis volcanic province, and the potential for the resulting generation of groundwater recharge through basal melting. This assessment demonstrated that even within volcanically active terrains, in the presence of regionally elevated geothermal heat fluxes, widespread basal melting would again require either anomalously large local ice thicknesses or combined top-down heating. However, over very local scales in close proximity to active volcanic features such as vents, calderas, or edifices, ice sheet basal melting could take place under nominal conditions leading to the so-called “heat pipe drain pipe” local aquifer recharge. Due to the limited areal extent over which the “heat pipe drain pipe” was found to be possible, the amount of aquifer recharge provided by the mechanism was determined to be minimal. As a result of the difficulties associated with

the generation of basal melting of martian ice sheets found through this earlier work, the focus of the subsequent work in this dissertation transitioned to top-down heating and melting.

In the first chapter of this dissertation, a new model is proposed to account for the generation of liquid water in an ambient cold and icy hydrologic system through large-scale, surficial lava-ice interactions. This model is proposed in light of recent climate model results predicting the formation of regional ice sheets in spatial and temporal coincidence with the emplacement of effusive flood basalt lavas that resurfaced $\sim 1/3$ of Mars during its early history. In this first chapter, fundamental model concepts are introduced and developed through a theoretical analysis of the proposed large-scale, surficial lava-ice interaction processes to produce quantitative estimates of meltwater generation and to outline predictions for resulting landforms/morphologies. This assessment is carried out through a case study of the Hesperia Planum region on Mars, the type area for the effusive Late Noachian – Early Hesperian ridged plains flood basalt unit. The chapter concludes with a preliminary test of the theoretical model by comparing case study predictions to the geomorphology of a suite of features in the Hesperia Planum region. The analyses of this chapter demonstrate the ice sheet lava heating and loading process to be a viable mechanism to produce significant ice melting, infiltration, runoff, surface modification, and to plausibly explain a suite of characteristic geological features of the Hesperia Planum study area. The fundamental concepts and model developed in this chapter are then improved upon and more rigorously tested in subsequent chapters through focused applied case studies of several martian hydrologic features.

The second chapter builds upon the groundwork developed in the first chapter through improvements to the theoretical treatment and a more rigorous test of the proposed large-scale, surficial lava-ice interaction model. Specifically, the model is tested as a potentially plausible origin for the martian outflow channels, enormous diluvial features carved into the surface of Mars throughout its history. The formation of the outflow channels of Mars has been previously attributed to the catastrophic effusion of groundwater. However, as is detailed in this chapter, the formation of these features by groundwater outbursts requires a number of exceptional circumstances and challenging conditions. The proposed large-scale, surficial lava-ice interaction model is not subject to these difficulties and therefore offers a potentially advantageous explanation for the formation of the martian outflow channels. The viability of large-scale, surficial lava-ice interactions as an origin for martian outflow channels is carried out in this chapter through an applied case study of Reull Vallis. The Reull Vallis outflow channel is a segmented system of fluvial valleys that emanates from the volcanic plains of the Hesperia Planum region of Mars, selected for the purposes of this evaluation due to its close association with the effusive volcanic plains of Hesperia Planum. The geomorphology of the Reull Vallis system is first characterized in order to outline criteria that must be met by the proposed formation mechanism. Regional lava emplacement conditions are then constrained and utilized to generate model predictions for the formation of Reull Vallis which are tested against the observational geomorphic criteria to assess consistency. The investigated ice sheet lava heating and loading model is found to offer a plausible explanation for the origin of the Reull Vallis outflow channel system

with several advantages over previous genetic models invoking the catastrophic effusion of groundwater.

In the third chapter, the lava-ice interaction model developed in the prior chapters is explored as a possible viable origin of the enigmatic outflow channels of Elysium Planitia. The outflow channels of Elysium Planitia are suggested by age dating estimates to have formed during the Amazonian period. These age estimates therefore indicate an anomalously young geologic age of the Elysium Planitia system relative to the broader population of outflow channels which date predominantly to the Noachian and Hesperian periods of Mars history. As a result of the young Amazonian age estimates for the Elysium Planitia outflow channels, accounting for the formation of these systems by the canonical groundwater outburst models faces a number of challenging factors which are reviewed in this chapter. Given these difficulties, this chapter revisits the origin of the Elysium Planitia outflow channels to explore large-scale lava-ice interactions as a possible alternative mechanism to contribute to, or account for, their formation. This assessment is performed through an applied case study of Athabasca Valles, the most intensely studied and best characterized outflow channel system of the region. As in the preceding chapter, the geomorphology of Athabasca Valles is first reviewed in order to outline required criteria that must be satisfied to account for formation of the channel. These criteria are then tested against predictions of meltwater generation and geomorphology from the earlier large-scale lava-ice interaction modeling work. Formation of the Athabasca Valles outflow channel system is found to have plausibly been the result of surficial lava-ice interactions with subsequent meltwater confinement, accumulation, and release at elevated discharge levels.

In the fourth chapter, the broad potential implications of the lava-ice interaction model are explored through an application of the model to perform a regional study of the Tharsis region, the largest volcanic province on Mars. The Tharsis region is an area predicted to have undergone surface ice accumulation under a wide range of climatic conditions throughout the history of Mars. In addition, throughout this same timeframe, the Tharsis region has been a site of intense and prolonged volcanic activity. Therefore, the Tharsis region has been a site of coincident volcanic and glacial activity throughout the majority of the geologic history of Mars. Given these prolonged and overlapping histories of volcanism and glaciation, the Tharsis region represents a prime candidate for an expanded assessment of the lava-ice interaction model. In this chapter, first-order, exploratory study is performed to assess the potential role that surficial lava-ice interactions may have played in the geologic and hydrologic history of the Tharsis region. We first review the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, in order to outline relevant conditions and parameters. The outlined volcanic and glacial conditions are then used in conjunction with results and predictions from the surficial lava-ice interaction model to assess the potential role these processes may have played in the formation of the Tharsis region's major tectonic and hydrologic features. Following a procedure similar to that of the preceding chapters, assessment of these features is carried out through a geomorphic characterization to derive necessary formation criteria which are then compared for consistency against predictions from possible surficial lava-ice interaction genetic models. Genetic models involving surficial glaciovolcanism are found to have possibly contributed to the formation of many of the tectonic and fluvial features

in the Tharsis region, offering advantages over prior formation models, particularly for the large basin/chaos-sourced outflow channels concentrated in the area

The fifth chapter focuses on the valley networks of Mars, systems of linear-to-sinuuous depressions which are widely distributed across the surface of the planet and which are interpreted to have formed by fluvial activity. Morphologic characteristics of the valley networks have often been interpreted to indicate the formation of these features through precipitation-derived fluvial activity and therefore as evidence for a relatively warm and wet Noachian Mars climate. This interpretation conflicts with the cold and icy conditions predicted by sophisticated global climate modeling studies. These contrasting predictions are addressed in this chapter through the development and evaluation of a conceptual model for valley network formation under cold and icy conditions to determine if the characteristics of the valley networks are consistent with formation in a cold and icy environment. The proposed conceptual model for valley network formation under cold and icy conditions is evaluated through the implementation of numerical thermal models, quantitative erosion and transport estimates, and morphometric analyses in order to outline predictions for resulting channel/valley morphology. Morphologic predictions generated through this analysis are then compared against observational data to assess the consistency of the valley network characteristics to those predicted of formation in a cold and icy environment. Through this approach, the underlying conceptual model is presented and developed, fundamental data inputs are identified and generated, and a preliminary first-order assessment is performed to serve as a basis for further investigation. Results of the theoretical analysis of the conceptual model suggest that the morphology of the valley networks could have been influenced by cold and icy climate

conditions. However, global examination of the valley network morphometry yielded inconclusive results which are attributed to a number of complicating factors and simplifying assumptions requiring further investigation.

In the sixth and final chapter, a synthesis of the preceding chapters is presented to frame the implications and importance of the developments of this dissertation toward understanding the climatic and hydrologic evolution of Mars. In addition to this synthesis, major outstanding questions and suggested paths of future investigation are outlined to continue to better resolve the nature of the hydrologic cycle of early Mars as well as the implications for habitability and the development of life.

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**Chapter 1: Lava Heating and Loading of Ice Sheets on Early Mars:
Predictions for Meltwater Generation, Groundwater Recharge, and
Resulting Landforms**

James P. Cassanelli^a & James W. Head^a

^a Department of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

Published in: *Icarus*

Cassanelli, J. P., and Head, J.W., (2016). Lava heating and loading of ice sheets on early Mars: Predictions for meltwater generation, groundwater recharge, and resulting landforms. *Icarus*, 271, 237-264, doi: 10.1016/j.icarus.2016.02.004

Abstract

Recent modeling studies of the early Mars climate predict a predominantly cold climate, characterized by the formation of regional ice sheets across the highland areas of Mars. Formation of the predicted “icy highlands” ice sheets is coincident with a peak in the volcanic flux of Mars involving the emplacement of the Late Noachian – Early Hesperian ridged plains unit. We explore the relationship between the predicted early Mars “icy highlands” ice sheets, and the extensive early flood volcanism to gain insight into the surface conditions prevalent during the Late Noachian to Early Hesperian transition period. Using Hesperia Planum as a type area, we develop an ice sheet lava heating and loading model. We quantitatively assess the thermal and melting processes involved in the lava heating and loading process following the chronological sequence of lava emplacement. We test a broad range of parameters to thoroughly constrain the lava heating and loading process and outline predictions for the formation of resulting geological features. We apply the theoretical model to a study area within the Hesperia Planum region and assess the observed geology against predictions derived from the ice sheet lava heating and loading model. Due to the highly cratered nature of the Noachian highlands terrain onto which the volcanic plains were emplaced, we predict highly asymmetrical lava loading conditions. Crater interiors are predicted to accumulate greater thicknesses of lava over more rapid timescales, while in the intercrater plains, lava accumulation occurs over longer timescales and does not reach great thicknesses. We find that top-down melting due to conductive heat transfer from supraglacial lava flows is generally limited when the emplaced lava flows are less than ~10 m thick, but is very significant at lava flow thicknesses of ~100 m or greater. We find that bottom-up

cryosphere and ice sheet melting is most likely to occur within crater interiors where lavas accumulate to a sufficient thickness to raise the ice-melting isotherm to the base of the superposed lavas. In these locations, if lava accumulation occurs rapidly, bottom-up melting of the ice sheet can continue, or begin, after lava accumulation has completed in a process we term “deferred melting.” Subsurface mass loss through melting of the buried ice sheets is predicted to cause substantial subsidence in the superposed lavas, leading to the formation of associated collapse features including fracture systems, depressions, surface faulting and folding, wrinkle-ridge formation, and chaos terrain. In addition, if meltwater generated from the lava heating and loading process becomes trapped at the lava flow margins due to the presence of impermeable confining units, large highly pressurized episodic flooding events could occur. Examination of the study area reveals geological features which are generally consistent with those predicted to form as a result of the ice sheet lava heating and loading process, suggesting the presence of surface snow and ice during the Late Noachian to Early Hesperian period.

1. Introduction

Global climate modeling studies of the early Mars climate (Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015) predict predominantly cold conditions under which liquid water is not stable at the surface of the planet. These predictions are generally inconsistent with a “warm and wet” early Mars climate (Craddock and Howard, 2002) interpreted from the widespread presence of valley network systems (Howard, 2007; Fassett and Head, 2008; Barnhart et al., 2009; Hynek et al., 2010; Hoke et al., 2011), open and closed-basin lakes (Cabrol and Grin, 1999; Carr, 2006; Fassett and Head, 2008b), and phyllosilicate-bearing units in Noachian-aged terrains (Bibring et al.,

2006; Ehlmann et al., 2011), as well as the degraded state of Noachian-aged impact craters (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Weiss and Head, 2015). Results from recent investigations modeling a thicker early Mars CO₂ atmosphere (Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2014), suggest that increased atmospheric pressure causes atmosphere-surface thermal coupling. This coupling leads to adiabatic cooling and a decrease in the mean annual temperatures of high elevation regions across Mars. The cooled highland regions act as cold traps and experience preferential accumulation of water-ice, leading to establishment of regional highland ice sheets which characterize the “icy highlands” early Mars climate model (Wordsworth et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2014; Head and Marchant, 2014).

The formation of regional ice sheets in the martian highlands predicted by the “icy highlands” model is coincident with the transition from the Late Noachian to Early Hesperian period on Mars, a time of dramatic change in both the geologic and climatic evolution of the planet (Carr and Head, 2010). The transitional Hesperian period of Mars history is characterized by a shift in the dominant mineralogic weathering style (Bibring et al., 2006), a sharp decrease in evidence for flowing surface water (*e.g.* valley networks; Fassett and Head, 2008), and a peak in volcanic flux (Craddock and Greeley, 2009; Carr and Head, 2010; Tanaka et al., 2014). The observed peak in volcanic activity occurred in the late Noachian and early during the Hesperian (Craddock and Greeley, 2009; Rogers and Nazarian, 2013; Tanaka et al., 2014a) in the form of planetary-scale flood volcanism, involving vast outpourings of volcanic material that resulted in the resurfacing of a significant portion of the planetary surface (Head et al., 2002; Head et al., 2006; Goudge

et al., 2012; Rogers and Nazarian, 2013). Conversely, fluvial activity waned throughout the Hesperian period (Fassett and Head, 2008a), followed by the emergence of the outflow channels during the Late Hesperian (Baker and Milton, 1974; Sharp and Malin, 1975; Carr, 1979; Carr and Clow, 1981; Baker, 1982; Baker et al., 1992; Carr, 1996; Carr, 2000; Baker, 2001; Burr et al., 2009; Irwin and Grant, 2009; Carr, 2012).

Understanding the individual nature of these processes can provide insight into the conditions of the martian interior as well as the climate and state of volatiles at the surface of the planet. In addition, understanding the relationship between these major surface processes may be able to provide further information into the nature of the conditions on the surface of Mars during the Late Noachian to Early Hesperian transition.

Here we explore the relationship between the predicted early Mars “icy highlands” ice sheets, and the extensive Late Noachian and Early Hesperian flood volcanism to gain insight into the surface conditions prevalent during this critical transition period. In the “icy highlands” climate scenario, accumulation of snowfall would lead to the deposition of regional ice sheets hundreds of meters thick within the Hesperia Planum region (Wordsworth et al., 2013; Head and Marchant, 2014; Wordsworth et al., 2015) which could then participate in a number of volcano-ice interactions including: (1) Direct interaction between extrusive lavas and surficial snow and ice deposits resulting in heating, melting, and explosive events (*e.g.* Wilson and Head, 2007). (2) Thick accumulations of lava could provide a thermal blanket, raising the ice-melting isotherm thereby inducing widespread melt of cryospheric ice leading to groundwater recharge (Clifford, 1993; Carr and Head, 2003; Russell and Head, 2007; Clifford et al., 2010; Cassanelli et al., 2015). (3) Accumulation of supraglacial lavas could raise the ice-

melting isotherm into surface ice deposits causing widespread basal melting (similar in nature to the effect of sedimentary materials superposed upon ice sheets as investigated by Zegers et al., (2010)). (4) Localized heating from individual volcanic structures (e.g. edifices, vents; Head and Wilson, 2002; Shean et al., 2005; Head and Wilson, 2007; Kadish et al., 2008; Scanlon et al., 2014) could serve as heat-pipes (Cassanelli et al., 2015), removing the cryosphere (defined as the portion of the martian crust which lies between the surface and the depth of the ice-melting isotherm) and inducing ice sheet basal melting at local scales. (5) Volcanic emissions (e.g. Craddock and Greeley, 2009) could produce episodic greenhouse atmospheric warming and allow transient top-down ice sheet melting events (e.g. Halevy and Head, 2014). These interactions are not predicted to occur in a “warm and wet” climate scenario, because under these conditions, a vertically integrated hydrological cycle (e.g. Craddock and Howard, 2002) would allow surface water to infiltrate into the subsurface, limiting interactions with extrusive volcanic processes. Therefore, evidence of surficial volcano-ice interactions during the Late Noachian to Early Hesperian transition would provide support for the dominance of “cold and icy” conditions at this time of Mars history.

To perform this assessment, we examine Hesperia Planum, the type area for the Hesperian Period (Figure 1) and a region of predicted “icy highlands” ice sheet formation which contains an array of volcanic and fluvial features and evidence for volatile-related processes (Squyres et al., 1987; Crown et al., 1992; Mest and Crown, 2001; Mest and Crown, 2002; Mest and Crown, 2003; Ivanov et al., 2005; Gregg and Crown, 2009; Mest and Crown, 2014). Hesperia Planum is an extensive $\sim 2 \times 10^6 \text{ km}^2$ Hesperian-aged smooth plains unit (Gregg and Crown, 2005) and the type location for the

Hesperian Ridged plains (Tanaka et al., 2014b), a unit characterized by the presence of wrinkle ridge structures and interpreted to represent effusive flood volcanic deposits (Head et al., 2002; Tanaka et al., 2014b). Previous contributions have explored the role of volcano-ice interactions in the Hesperia Planum region (Squyres et al., 1987) and have invoked the interaction of intrusive and extrusive volcanism with ground-ice in the formation of several observed features. Here, in light of the predictions made by the recent “icy highlands” model, we re-examine the role of volcano-ice interactions in the Hesperia Planum region and test an “ice sheet lava heating and loading” mechanism in which the Early Hesperian volcanic plains are emplaced atop the predicted regional “icy highlands” ice sheets.

In this contribution we detail a theoretical analysis of the lava heating and loading mechanism treating each aspect of the process in chronological sequence. (1) We first examine the volcano-ice interactions and melting processes associated with the emplacement of an initial lava flow, which we define as a lava flow which is emplaced directly upon the ice sheet surface. (2) We then modify and extend this initial treatment to assess the effects and melting resulting from emplacement of subsequent lava flows. (3) Lastly, we outline and implement a numerical model to test the long-term effects and bottom-up melting processes resulting from continued ice sheet lava heating and loading. The analyses presented here assume ice sheet formation prior to the onset of volcanic activity. However, accumulation of ice and lava could have been coeval, resulting in interspersed deposits of ice and lava. The implications of coeval ice and lava accumulation for the interactions and processes we explore are discussed in a later section.

Results from these analyses are used to synthesize predictions for the generation of geological features, which are then compared to the geological record observed in a study region within Hesperia Planum. This morphological comparison is used to derive implications for the prevailing conditions during the Late Noachian to Early Hesperian transition period.

2. Lava Thicknesses and Accumulation Timescales

The total thickness of lava accumulated atop the ice sheet, along with the thickness of individually emplaced lava flows and the accumulation timescale are the most important factors controlling the thermal aspects of the lava heating and loading process. This is because these factors determine the amount and timing of heating and insulation being provided to the loaded ice sheet. Therefore, we begin by making estimates of the range of total volcanic plains lava thicknesses in the Hesperia Planum study region. In order to make this determination, diameters and topographic measurements were collected from 51 buried and partially buried impact craters (Figure 2; SI Table 1). We first make the assumption that the measured crater diameters are representative of the fresh crater diameters and then translate the measured crater diameters to post-collapse crater rim height and crater depths through crater scaling laws (Craddock et al., 1997; Tornabene et al., 2014). The crater rim crest heights and diameters calculated in this manner serve as upper limits because many of the measured buried craters may have been degraded prior to lava flow emplacement, which would reduce both the crater rim crest height and crater depths. Noting that these estimates will reflect upper limits, we obtain minimum accumulated lava thicknesses by assuming that lava must have accumulated to at least the height of the crater rim crests to bury the crater. Maximum accumulated lava thicknesses

were then obtained by assuming that all the filling material inside the crater is sourced from the Hesperian volcanic plains (thus the difference between the currently observed crater floor depth, and the fresh crater depth give the maximum accumulated lava thickness). This estimate also provides an upper limit since the measured craters may have been filled to some extent by other materials prior to lava emplacement. Histograms showing the distribution of measurements taken from the Hesperia Planum are shown in Figure 2 and the measured crater locations and diameters are tabulated in SI Table 1. Based on these assumptions the collected measurements yield an average minimum accumulated lava thickness of ~500 m, and an average maximum accumulated lava thickness of ~2 km.

To estimate the thickness of the individual lava flows we consider the general nature of the Hesperia Planum volcanic plains which are thought to have been emplaced in a flood basalt mode (Head and Coffin, 1997; Head et al., 2002; Head et al., 2006), similar in nature to the terrestrial continental large igneous provinces (*e.g.* Coffin and Eldholm, 1994; Self et al., 1997). In the terrestrial continental large igneous provinces, lava flows are emplaced (Reidel et al., 2013) as either distinct individual series of compound flows, or as extensive, high-volume sheet flows. In general, the continental flood basalt provinces are predominantly constructed from the accumulation of sheet flows (Reidel et al., 2013). The individual thickness of these sheet flows varies considerably from ~1 m to as much as 150 m (Self et al., 1997; Sharma, 1997), with flow thicknesses observed in the Columbia River Flood basalt province generally in the range of ~10-50 m (Self et al., 1997). These lava flow thicknesses bracket the observed terrestrial flood basalt flow thicknesses and form reasonable guidelines for the flow thicknesses expected in the

Hesperia Planum volcanic plains. However, due to the gravity scaling of the Bingham rheology of lava (Hulme, 1974), flows on Mars would be ~ 2.5 times as thick as terrestrial flows, all else being equal. In addition, the Hesperia Planum volcanic plains were emplaced upon a heavily cratered Noachian-aged highlands unit which is likely to have complicated the emplacement and accumulation of the flood basalt plains (Head, 1982; Whitten and Head, 2013) relative to the terrestrial case. Modeling the volcanic flooding of heavily cratered planetary surfaces by Whitten and Head (2013) has shown that impact craters act as focal points for lava accumulation. This causes more rapid local lava flooding and an effective increase in lava flow thicknesses within crater interiors relative to the surrounding intercrater plains. As a result, a dichotomy in lava emplacement conditions between crater interiors and the intercrater plains is predicted in the Hesperia Planum region. This will result in asymmetrical emplaced lava flow thicknesses at both local and regional scales within Hesperia Planum. To address the effects of martian gravity and the predicted asymmetrical nature of individual lava flow thicknesses within the Hesperia Planum region, we test a broad range of lava flow thicknesses from 1-200 m (which encompasses the observed range of lava flow thicknesses in terrestrial continental large igneous provinces).

Previous estimates on the timescale of Hesperian ridged plains emplacement suggest a total emplacement time of ~ 100 -200 Myr (Craddock and Greeley, 2009, Tanaka et al., 2014a) with active eruption periods occupying only $\sim 0.01\%$ of this time, giving a total cumulative eruption duration of ~ 10 kyr (Halevy and Head, 2014). The proportion of the total emplacement timescale occupied by active eruption periods interpreted for the Hesperian ridged plains is consistent with the terrestrial Columbia River flood basalt

large igneous province which shares a similar relationship between the total emplacement timescale and cumulative eruptive duration (Self et al., 1997; Reidel et al., 2013). To assess the range of possible lava accumulation timescales, we test end-member cases whereby lava emplacement is completed in a minimum of 10 kyr and a maximum of 100 Myr (chosen over a 200 Myr case to reduce the required computational time).

3. Initial Lava Flow Emplacement

Quantitative assessment of the ice sheet lava heating and loading mechanism begins by considering the emplacement of the initial lava flow, which we define as a lava flow which is emplaced directly upon the ice sheet surface. The emplacement of lava flows atop an ice sheet can occur by two main mechanisms. Lava flows can be emplaced across the ice sheet surface by simply advancing onto the top of the ice sheet from a topographically elevated non-ice covered location (as is observed to occur in some terrestrial volcanic settings; Edwards et al., 2015), or by dike emplacement through the ice sheet due to high strain rates (Wilson and Head, 2002; Head and Wilson, 2002; Head and Wilson, 2007), leading to the eruption of lava flows at the ice sheet surface.

3.1 Thermal Analysis

Following emplacement, the initial lava flow will begin to undergo conductive cooling, transferring heat into the underlying ice sheet, and to the atmosphere above. To determine the amount of heat that is transferred to the underlying ice, we solve the one-dimensional heat conduction equation $\left(\frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2}\right)$ following Wilson and Head (2007). We treat the lava flow as an infinite slab (considering heat transfer in only the vertical direction), and assume that the flow is emplaced instantaneously relative to the duration of heat transfer and cooling (which is the case for most lava flows; Pinkerton and Wilson,

1994). For the thinner lava flows (1 m and 10 m in thickness) we apply an initial sinusoidal temperature distribution throughout the lava slab to account for a lava flow structure exhibiting chilled flow margins, with temperatures increasing towards the lava flow core reaching a peak value of approximately 1350 K (as observed in the supraglacial flows of the 2012-2013 Tolbachik eruption on the Kamchatka peninsula in Russia; Edwards et al., 2015). For the lava flows of greater thickness (100 m and 200 m), we assume that the chilled lava flow margins will have little effect on the overall temperature distribution of the lava flow structure and apply a uniform temperature throughout the lava slab. For this uniform lava flow temperature we take an average of lava flow temperatures derived from geothermometer measurements of the Columbia River Flood basalts (Self et al., 1997), giving a value of 1350 K. The temperature at the top of the slab in all cases is held constant at a predicted Late Noachian mean annual surface temperature (225 K; Wordsworth et al., 2013; Wordsworth et al., 2015) and the temperature at the base of the slab is held constant at the melting point of water (assuming a continuous interface with ice at the base of the lava flow is maintained). Solution of the heat equation subject to these conditions over the thickness of the lava flow (in the z -direction on the interval $0 < z < L$) produces the following series:

$$T(z, t) = T_S + (T_B - T_S) \frac{z}{L} + \sum_{j=1}^n A_j \sin(j\pi z/L) e^{-k j^2 \pi^2 t/L^2} \quad (1)$$

where:

$T(z, t)$ = temperature (K) at any depth z (m) within the flow at time t (s)

T_S = surface Temperature (K)

T_B = basal Temperature (K)

A_j = fourier coefficient for the initial temperature distribution

L = lava flow thickness (m)

k = thermal diffusivity (m^2/s)

We compute the series given by equation (1) to $n = 50$ terms at each time value evaluated to ensure solution convergence even at small values of time t (~ 600 s). The heat flux into the ice beneath the lava flow is: $H(W/m^2) = K_T \frac{dT}{dz}$ with the thermal conductivity $K_T = k\bar{\rho}c$ where k is the thermal diffusivity, $\bar{\rho}$ is the bulk density of the lava (kg/m^3), c is the specific heat capacity ($J/kg\ K$), and the temperature gradient $\left(\frac{dT}{dz}\right)$ at the lava-ice interface is calculated by extrapolating the gradient from $0.98 \times z$ to $0.999 \times z$. The rate of melting of the underlying ice sheet is then $R(m/s) = H / \rho L_i$ where L_i is the latent heat of fusion of ice ($\sim 3.35 \times 10^5 J/kg$), and ρ is the density of the underlying ice sheet ($917 kg/m^3$ for solid ice). The total thickness of ice melted following lava flow emplacement is calculated by integration of the melting rates throughout the entirety of the cooling process. We note that the melting rates and total ice thicknesses melted that are predicted by this analysis are effected by several assumptions implicit in the analysis. These include the assumption of constant thermal properties for the cooling basalt, the assumption that no heat energy is expended towards warming the water produced after melting has occurred, and disregarding the energy released from the latent heat of fusion when the lava flow cools below the solidus temperature. The effects of these assumptions on the thermal analysis are not predicted to be substantial and are quantified and discussed in Wilson and Head (2007).

We assume a basaltic magma composition (typical of Hesperian ridged plains volcanic materials as observed within Gusev crater; McSween et al., 2006), and take typical terrestrial values for the basaltic lava density ($\sim 3,000 kg/m^3$), specific heat capacity ($\sim 900 J/kg\ K$), and thermal diffusivity ($7 \times 10^{-7} m^2/s$) (Wilson and Head, 2007).

The heat transfer rates into the underlying ice are calculated throughout the entirety of the lava flow cooling period with the use of equation (1) for each of the adopted lava flow thicknesses discussed in section 2. The heat transfer rates are then integrated throughout the cooling period of each lava flow to determine the amount of ice that is melted versus time following lava flow emplacement (Figure 3).

3.2 Initial Lava Flow Emplacement: Snow and Firn Layer

It has been assumed to this point that supraglacial lava flows emplaced atop the surface of an ice sheet will be in contact with a surface of pure ice. However, in many cases the surficial layers of ice sheets are comprised of snow and firn instead of pure ice (Cuffey and Paterson, 2010) (where snow is defined as water-ice having a bulk density of less than 360 kg/m^3 , firn as water-ice with bulk densities from $360\text{-}830 \text{ kg/m}^3$, and ice for bulk densities greater than 830 kg/m^3). The presence of a snow and firn layer at the ice sheet surface will enhance the melting rates and total thicknesses melted during supraglacial lava flow emplacement due to the reduced bulk densities of the snow and firn relative to solid ice.

The snow and firn layer is a result of the construction of the ice sheet from accumulating snow which transitions into ice through a firn densification process (*e.g.* Cuffey and Paterson, 2010; Cassanelli and Head, 2015). On Earth, ice sheet firn layers are found primarily in the cold accumulation zones of ice sheets (Cuffey and Paterson, 2010). This is because in the absence of accumulation, the firn continuously undergoes densification without replenishment and because in the ablation zones, temperatures above the melting point rapidly diminish the firn layer. On Mars, the low gravity and low surface temperatures favor the preservation of the firn layer, thus firn layers comparable

or greater in thickness to those found in the accumulation zone of terrestrial ice sheets are predicted to exist across the surface of any ice sheets present on Mars (Cassanelli and Head, 2015).

In the “icy highlands” scenario, ice sheet growth is predicted to be a supply-limited process (Carr and Head, 2015; Fastook and Head, 2015), resulting in the termination of snow accumulation once the water-ice supply has been exhausted. Without melting and recycling of the water stored in the “icy highlands” ice sheets establishing an equilibrium state, the thickness and density state of the ice sheet snow and firn layer will vary as a function of time during the ice sheet formation process (modulated by the prevailing climate conditions) (Cassanelli and Head, 2015). Under nominal “icy highlands” conditions the firn layer will be at a maximum thickness of ~ 115 m immediately after the completion of ice sheet formation once the water ice supply has been exceeded (Cassanelli and Head, 2015). The thickness of the firn layer will then reduce over time, reaching a thickness of ~ 17 m after 1 Myr of ice sheet evolution and densification without further snow accumulation (Cassanelli and Head, 2015). The presence of a firn layer has an appreciable effect on the results of the thermal analysis which we now discuss.

3.3 Initial Lava Flow Emplacement: Results

We find that following emplacement, thinner lava flows contribute an initially higher heat flux to the underlying ice due to more efficient cooling. As a result of the higher heat transfer rates and the lower total heat energy that is contained within thinner lava flows, thinner lava flows cool to ambient temperatures significantly more quickly than the thicker lava flows. Therefore, despite the higher initial heat transfer rates provided by

thinner lava flows, the sustained nature of heat delivery from thicker lava flows due to prolonged cooling and the increased total heat energy, ultimately results in the melting of a much greater thickness of ice (Figure 3).

The cooling timescales of initial lava flows calculated through the thermal analysis range from ~7 days to ~800 years for the evaluated lava flow thicknesses of 1-200 m (Figure 3). Integration of the heat transfer rates throughout the cooling period for each lava flow indicates a near constant ratio between the thickness of ice melted and the thickness of the initial lava flow emplaced equal to ~3 for the thinner lava flows (1 m and 10 m in thickness) and ~4.5 for the thicker lava flows (100 m and 200 m in thickness) (Figure 3). The difference in this ratio between the thinner lava flows and thicker lava flows is due to the different initial temperature distributions applied within the thinner and thicker lava slabs, however, both results are in general agreement with results determined for the terrestrial case from Wilson and Head (2007).

Thus, throughout the initial lava flow cooling process, a 1 m lava flow will transfer enough heat to melt ~3 m of ice, while a 10 m lava flow will be able to melt ~30 m of ice (Figure 3). For the thicker lava flows, the thermal analysis results suggest that the 100 m, and 200 m thick initial lava flows contain enough heat energy to melt through a 450 m and 900 m thick ice sheet, respectively (Figure 3). The implications of this significant top-down melting potential are discussed in more detail in later sections. In all cases, the lava flows are predicted to undergo subsidence equal to the total thickness of ice melted, assuming efficient evacuation of meltwater.

The total thicknesses of ice melted during the emplacement and cooling of each evaluated lava flow thickness are shown in Table 1 along with the total melted thickness

adjusted for the presence of the end-member firn layer thicknesses. These results indicate: (1) The 1 m thick initial lava flow will not be able to melt completely through either modeled firn layer. (2) The 10 m lava flow will be able to melt through only the thin firn layer. (3) The melting totals associated with the thicker lava flows are little affected by the presence of the firn layer. (4) The amount of subsidence each lava flow is predicted to undergo during the top-down melting process is increased due to the presence of the firn layer.

A fundamental change in the scale of predicted top-down melting exists between the thinner initial lava flows (1 m and 10 m) and the thicker initial lava flows (100 m and 200 m) evaluated here. The total melted thicknesses associated with the thinner lava flows do not account for a substantial amount of the entire predicted “icy highlands” ice sheet thicknesses (300-1,000 m; Carr and Head, 2015; Fastook and Head, 2015). Conversely, the total melted thicknesses predicted for the thicker lava flows approach, and even exceed, the thicknesses expected of the “icy highlands” ice sheets. Therefore, a fundamentally different regime of meltwater transport and fate processes will arise during the emplacement of the thicker lava flows. We now examine in detail the transport and fate of meltwater generated through top-down melting following lava flow emplacement, considering first the cases in which thinner initial lava flows are emplaced.

3.4 Top-down Meltwater Transport and Fate: Thin Lava Flows

Meltwater produced at the surface of the ice sheet during the emplacement and cooling of an initial lava flow can follow one of several pathways (Figure 4): (1) The meltwater may enter into storage within the underlying porous firn layer, whereby the firn layer acts much like a terrestrial groundwater aquifer (Forster et al., 2014). (2) The

meltwater may drain towards the glacial margins across the top of the ice sheet, forming channels as it follows ice sheet surface topography. (3) The meltwater may drain towards the ice sheet base through cracks, crevasses, or moulins. (4) Meltwater may pool beneath the lava flow and at the lava flow margins, enhancing cooling and resulting in phreatomagmatic events, or refreezing after cooling has finished and temperatures have decreased.

The nature of the ice sheet snow and firn layer, and the thickness of the emplaced lava flow will determine which of these meltwater pathways is dominant (Figure 4). Under the Late Noachian conditions of interest, there are two broad possibilities for the state of the ice sheet surface: (1) The firn layer may be relatively thick if lava emplacement has occurred shortly after the completion of ice sheet formation or if some melting mechanism allows for water recycling and continued snow deposition. (2) The firn layer may be relatively thin if the ice sheets have remained stable over timescales on the order of ~ 1 Myr after formation without the input of additional snow to replenish the firn layer.

In either of these scenarios, the porous snow and firn layer will act to subdue runoff across the surface of the ice sheet by absorbing meltwater, thereby eliminating this meltwater pathway. The absorption of meltwater from the melting interface with the lava flow will also reduce the likelihood of phreatomagmatic eruptions by evacuating water from the lava interface and suppressing steam generation. Meltwater may drain towards the ice sheet base through cracks, crevasses, and moulins. However, due to the predicted pervasiveness of snow and firn across the “icy highlands” ice sheets (Cassanelli and Head, 2015), the primary fate of water produced by top-down melting is predicted to be

absorption within the snow and firn layer. If meltwater production rates exceed the infiltration capacity of the porous snow and firn layer, then other interactions are possible, including localized meltwater ponding. Thus, estimating the infiltration capacity of the snow and firn layer is important.

To estimate the infiltration capacity of the firn layer we implement the following adaptation of Darcy's law for saturated groundwater flow (Hendriks, 2010):

$$I_R = K \left(\frac{L + S_f + h_o}{L} \right) \quad (2)$$

where:

I_R = infiltration rate (m/s)

K = hydraulic conductivity (m/s)

L = thickness of porous medium considered (m)

S_f = wetting front soil suction head (m)

h_o = head of infiltrating water ponded at porous media interface (m)

In order to estimate a minimum infiltration rate, we conservatively assume that both the wetting front soil suction head and the head of ponded infiltrating water are equal to zero. This reduces the infiltration rate predicted by equation (2) to the saturated hydraulic conductivity of the substrate. While this implementation of Darcy's law is valid for the flow of water through an unsaturated media, the unsaturated hydraulic conductivity is not a constant, but is a function of the porous medium water content (Fetter, 2001). The hydraulic conductivity will be at a minimum when the water content is zero, and will increase to a maximum at saturation. In practice the relation between water content and the unsaturated hydraulic conductivity is determined experimentally (typically varying by about an order of magnitude; Fetter, 2001). Here we adopt the saturated hydraulic conductivity, and note that the calculated infiltration rates will serve as upper limits.

The hydraulic conductivity K_h (m/s) of the snow and firn can be calculated from the intrinsic permeability $k(m^2)$ as:

$$K_h = \frac{k\rho_w g}{\mu} \quad (3)$$

where ρ_w is the density of water (1000 kg/m^3), μ is the dynamic viscosity of water ($1.793 \times 10^{-3} \text{ Pa} \cdot \text{s}$ at 273 K), and g is the gravitational acceleration in m/s^2 . Colbeck and Anderson (1982) measured the saturated permeability of melting snow, at a density of 400 kg/m^3 , to be $\sim 4 \times 10^{-9} m^2$. Under martian gravity this permeability gives a saturated hydraulic conductivity of $\sim 30 \text{ m/hr}$. This infiltration rate is more than two orders of magnitude greater than the highest meltwater production rate predicted from lava flow emplacement, which is $\sim 100 \text{ mm/hr}$, achieved during the initial supraglacial emplacement of a 1 m thick initial lava flow. Therefore, the infiltration capacity of the snow and firn layer is likely more than sufficient to accommodate the predicted meltwater production rates (even if a significant reduction in conductivity due to unsaturated conditions is taken into consideration). As a result, meltwater generated at the lava-ice interface will percolate downwards into the snow and firn layer, where it will be subject to eventual refreezing. When the meltwater refreezes, latent heat of fusion will be released from the melt into the surrounding ice or firn at depth which will enhance the densification of the surrounding material firn, aiding in the transition to ice.

Any snow and firn that remains after lava flow cooling has completed will continue to undergo compaction and densification (Cassanelli and Head, 2015) in response to the overburden stress of the emplaced lava flows. Firn that is compacted into ice will result in a reduction of the firn layer thickness by an amount:

$$z * \left(1 - \frac{\int_0^z \rho(z) dz}{\rho_i} \right) \quad (4)$$

where, z is the thickness of the firn layer compacted, $\rho(z)$ is the density of the firn layer as a function of depth (z) prior to compaction, and ρ_i is the density of ice. However, this component of firn reduction is likely to be negligible because the lava flows are very effective at removing the firn layer and because the highly porous near surface firn layers most susceptible to compaction will have been removed by melting.

As the firn layer undergoes densification and becomes diminished through melting, compaction, and refreezing, the porosity and permeability will decrease (causing the hydraulic conductivity to decrease to ~ 7 m/hr at a density of ~ 550 kg/m³, down to zero at a density of ~ 830 kg/m³ when impermeability is reached). The reduction in firn layer thickness and porosity will also cause a decrease in the storage capacity of the affected firn layer subjacent to the lava flow by reducing the pore space that can be occupied by meltwater. Prior to melting and compaction, the thin (17 m) firn layer is able to store a ~ 7 m column of meltwater per unit area, while the thick (115 m) firn layer is able to store a ~ 36 m column of meltwater per unit area. Taking into consideration firn layer reduction due to melting during lava flow emplacement and cooling, even the thinnest initial lava flow considered in this study (1 m) will produce more meltwater throughout the duration of cooling than the thin firn layer can accommodate, while a 10 m thick initial lava flow produces more water than the thick firn layer can accommodate.

Therefore, we predict that in most cases, more meltwater will be produced during initial lava flow emplacement than can be stored within the underlying firn. In cases where the firn layer is not completely removed, but the storage capacity is exceeded,

meltwater will migrate laterally into the unaffected surrounding firn (Figure 4). However, if the lava flow is able to melt completely through the firn and down into the impermeable ice, meltwater will no longer be able to migrate downwards away from the lava. As a result, meltwater that is not able to drain through cracks, crevasses, or moulins will pool beneath the lava flow. The weight of the overlying lava flow will impart hydraulic head to this meltwater, which will then direct the meltwater towards the lava flow margins (Figure 4). Once the lava flow margin has been reached, the meltwater may move upwards along the interface between the ice and the lava flow. If meltwater is not able to infiltrate into the surrounding firn layer as it rises along the lava flow margins (which could be prevented by low permeability surrounding firn or by ice) the lava flow may become flooded and inundated. This will accelerate the cooling of the lava flow, and may trigger phreatomagmatic events, though it will also result in less ice melting, since a portion of the heat energy of the lava flow will go towards heating the meltwater, and to production of steam. Phreatomagmatic eruptions will result in at least local destruction of the lava flows and firn layer, and dispersal of the lava flow material across the surrounding ice sheet surface. In addition a crater would form which would serve as a collecting location for meltwater and debris, which would then simply refreeze. If no eruption takes place, after cooling has finished, the water will freeze, potentially encapsulating the lava flow in ice.

3.5 Thin Initial Lava Flow Emplacement: Synthesis and Predictions

- The dominant transport pathways for meltwater produced by top-down melting following initial lava flow emplacement are downward percolation through the

porous firn layer, and drainage through any ice sheet fractures, crevasses, and moulins.

- Meltwater that is absorbed by the snow and firn layer will enter storage within the firn where it will refreeze (though the firn layer may offer temporary protection against refreezing; Forster et al., 2014). Refreezing will release latent heat of fusion energy into the surrounding ice or firn, enhancing the transition of firn to ice if refreezing occurs within the firn layer.
- Meltwater that intersects a fracture, crevasse, or moulin will be transported towards the ice sheet base, and refreeze at some depth within the ice sheet since the ice sheet will remain in a cold-based state (Fastook and Head, 2015).
- The expansion resulting from refreezing of the meltwater at depth within the ice sheet may result in the formation or extension of fractures within the ice sheet.
- We predict that supraglacial lava flows will typically be emplaced upon a relatively thin firn layer (~17 m) because under nominal Late Noachian conditions, the firn layer at the ice sheet surface will rapidly thin over geologically short time scales (~100 kyr; Cassanelli and Head, 2015).
- Emplacement of even the thinnest initial lava flows considered (~1 m) will effectively remove the thin predicted firn layer.
- Meltwater produced by thicker lava flows (~10 m) will only enter storage in the firn initially since the flow will melt completely through the firn layer and down into the solid ice. In this case meltwater will pool around the flow enhancing cooling and potentially resulting in phreatomagmatic events. After the lava flow has completed

cooling, the meltwater will refreeze around the lava flow, possibly encapsulating the flow if enough melt is pooled.

- The net effect of initial lava flow emplacement will be efficient removal of the surficial snow and firn layer, resulting in the subsidence, deformation, and degradation of the lava flow. The final degraded initial lava flows will then construct a cap across the ice sheet surface.

3.6 Top-down Meltwater Transport and Fate: Very Thick Lava Flows

Assessment of heat transfer and melting rates induced by the supraglacial emplacement of very thick lava flows (100 m and 200 m) suggests that these thick lava flows contain enough heat to melt thicknesses of ice which approach, or exceed, likely “icy highlands” ice sheet thicknesses (300-1,000 m; Carr and Head, 2015; Fastook and Head, 2015). As a result of the significant melting predicted to occur (Figure 3), the dominant meltwater pathways will now be (Figure 5): (1) Local pooling at the margins of the lava flow or ponding beneath the flow, (2) drainage to the ice sheet base through cracks, crevasses, or moulins, and potential lateral transport, and (3) drainage towards the glacial margins through surface channelization following the ice sheet surface topography.

In the case previously examined involving the supraglacial emplacement of thin lava flows (~1-10 m), the surficial firn layer was predicted to subdue ice sheet surface runoff through absorption of meltwater. However, in the case we examine now, where the emplaced lava thickness is much greater, the lava flow will rapidly melt down through the firn layer into the impermeable ice sheet where most of the melting will take place. As a result, meltwater that is generated by the lava flow will not be able to infiltrate away

from the lava flow unless it intersects an ice sheet fracture or moulin that allows drainage toward the ice sheet base. Instead the meltwater will be directed along the lava-ice interface toward the lava flow margins where it will ascend along the lava-ice contact under the influence of the overburden pressure of the lava flow (Figure 5). This meltwater will pool around and above the lava flow and will begin to absorb into firn that remains at the ice sheet surface surrounding the lava flow (Figure 5), though the surrounding firn may become overwhelmed by the quantity of water produced. In topographically favorable locations, meltwater is predicted to concentrate and to begin channelization across the ice sheet surface, eroding the firn layer as it drains towards the ice sheet margins (Figure 5).

After cooling has completed, any remaining meltwater pooled around the lava flow will refreeze in place. Similarly, meltwater that drains towards the ice sheet base will refreeze at depth within the ice sheet since the ice sheet will remain cold-based throughout the top-down melting process (Figure 5). This is because the insulation provided by the thick lava flows is not sufficient to raise the ice-melting isotherm to the ice sheet base, and because the top-down melting produced by the lava flows occurs over a much more rapid timescale (several 100 to several 10^4 years) than the bottom-up cryospheric heating and melting by raising the geotherm (many 10^6 years).

3.7 Very Thick Initial Lava Flow Emplacement: Synthesis and Predictions

- Predicted melting totals resulting from supraglacial emplacement of very thick lava flows can approach, and exceed, the total thicknesses of the “icy highlands” ice sheets.

- As a result of the significant melting, the firn layer has little effect on meltwater transport and fate.
- Meltwater is predicted to pool around the flow, drain to the ice sheet base, or runoff across the ice sheet surface in channels following surface topography.
- Meltwater that pools around the flow or drains to the ice sheet base will refreeze after lava flow cooling, while meltwater that drains to the glacial margins may form channels in the martian surface after it migrates beyond the ice sheet margins.
- The cryosphere underlying the ice sheet will remain intact during and after the initial thick lava flow emplacement because the insulation provided is not sufficient to dramatically raise the ice-melting isotherm.
- The net result of thick initial lava flow emplacement will be either considerable reduction in the total ice sheet thickness or complete top-down melting of the ice sheet (over a timescale of several 100 to 10^4 years; Figure 3).

4. Subsequent Lava Flow Emplacement

Following the emplacement and cooling of the initial supraglacial lava flow, the emplacement of subsequent lava flows will contribute much less heat to the underlying ice due to the intervening cooled initial lava flow. To estimate the heat flux delivered to the underlying ice sheet during subsequent lava flow emplacement, we performed the same thermal analysis used to assess the heat transfer from the initial lava flow with a modification of equation (1) to account for the different initial conditions. We assume that a subsequent flow, equal in thickness to the initial lava flow, is emplaced, thereby doubling the value of L in equation (1). We then apply the same initial distributions of heat, but across only the top half of the now larger modeled lava sequence (over the depth

region $0 < z < L/2$ where z is depth). The same boundary conditions are applied at the surface of the modeled lava sequence (at $z=0$) as well as the base (at $z=L$), based on the assumption that the second flow is emplaced after the cooling period of the initial lava flow. The remainder of the analysis is performed as defined in the section 3.1. The same analysis is repeated to model the emplacement of a third flow, by tripling the value of L in equation (1), and applying the initial heat distribution to the top third of the modeled lava sequence (over the depth region $0 < z < L/3$ where z is depth).

We find that the heat transfer from subsequent lava flows follows similar relationships with respect to lava flow thickness as for the initial lava flow, such that thinner flows produce higher, but less sustained, heat transfer rates relative to thicker flows (Figure 3). However, the onset of heat transfer from subsequent lava flows to the underlying ice is delayed from the emplacement time and reduced in magnitude relative to the initial lava flow, with peak heat flows occurring later in the cooling period (Figure 3). As a result, the meltwater production rates, and total thicknesses of ice melted, are substantially less than those for the initial lava flows which are emplaced directly upon the ice. The heat transfer analysis performed here indicates that for each successively emplaced subsequent lava flow, there is a consistent decay in the ratio of the total thickness of ice melted to the lava flow thickness. The decay in the total ice thickness melted, M_T (m), versus subsequent lava flow thickness can be approximated by the following relationship:

$$M_T = M_R * z / ((z + L) / z) \quad (5)$$

where M_R is the ratio of the total thickness of ice melted to the lava flow thickness (which has a value of ~ 3 for the thinner lava flows, and ~ 4.5 for the very thick lava flows), z is the thickness of the subsequent lava flow (m), and L is the total thickness of underlying cooled lava flows (m). Results showing the amount of ice melted versus time following subsequent lava flow emplacement for the evaluated range of lava flow thicknesses are shown graphically in Figure 3. With respect to the subsequent emplacement of the thin lava flows, the reduction in the total thickness of ice melted, the reduction in firn layer thickness, and the presence of the previously emplaced lavas, will result in a different series of meltwater transport and fate processes. Conversely, further melting is only predicted to occur after the subsequent emplacement of thick lava flows if the underlying ice sheets are very thick ($> \sim 500$ m), otherwise the ice sheet will have been completely melted by the initial lava flow. Due to the differences in melting conditions associated with subsequent lava flow emplacement, a different suite of meltwater transport and fate processes are predicted, which we now assess.

4.1 Subsequent Lava Flow Emplacement: Meltwater Transport and Fate

During the subsequent emplacement of thinner lava flows (1 m and 10 m thick), the potential transport pathways for meltwater produced from subsequent lava flow emplacement are effectively the same as those for the initial lava flow. The only significant difference is that in the case of the subsequent lava flow emplacement, the melting interface will exist at the base of the underlying initial lava flow (Figure 6). This interface will now initially lie at some depth below the surface of the ice sheet due to the subsidence of the initial flow caused by melting. As a result, the transport and fate of the

meltwater depends upon the state of the remaining firn layer beneath the initial lava flow, and upon the thickness of the lava flows (Figure 6).

If the firn layer was initially thick, then a non-trivial thickness of firn might remain after initial lava flow emplacement (*e.g.* ~60 m will remain if a 10 m lava flow was initially emplaced upon ~115 m of firn, disregarding any firn compaction that may have taken place in the intervening time). In this case, melt produced by the subsequent flows will be absorbed by the remaining firn (Figure 6) despite the diminished permeability since subsequent lava flows produce significantly lower heat fluxes, and melt much less ice (Figure 3). Following absorption, the meltwater will participate in the same processes outlined in section 3.5.

If the firn layer was initially thin it will most likely have been significantly reduced, or completely removed, by the emplacement of the initial lava flows. In this case, meltwater will not be able to move downwards due to the impermeable underlying ice and will instead migrate toward the lava flow margins and begin to ascend along the lava-ice contact (Figure 6) if no other path is available (*e.g.* fractures within the underlying ice). The meltwater can then become absorbed by the surrounding firn if it is encountered before the water ascends to the top of the lava flow sequence (Figure 6), or pool above the flow, submerging the subsequent lava flow, and increasing cooling rates, and the likelihood for phreatomagmatic events. Lava flow submergence is less likely to occur during subsequent flow emplacement, since much less melt is produced (Figure 3) due to the reduced heat transfer rates, and because the total lava flow sequence thickness is greater. In either case, the fate of the meltwater will be to undergo freezing within the

surrounding firm, or around the margins of the lava flow itself (Figure 6) as described in previous sections.

If the “icy highlands” ice sheets were sufficiently thick ($> \sim 500$ m), then the ice sheet may not have been completely melted during the emplacement of an initial 100 m thick lava flow (though it will have been nearly entirely removed by a lava flow 200 m in thickness). Therefore ice will remain to undergo melting during the subsequent emplacement of 100 m thick lava flows. However, continued accumulation of very thick flows will result in complete top-down melting of the predicted thicknesses of “icy highlands” ice sheets (~ 300 -1,000 m) during the emplacement of only ~ 1 -5 successive lava flows (Figure 3). Figure 7 depicts a synthesized illustration of the lava loading and heating process and meltwater transport and fate pathways in the case of successive emplacement and accumulation of very thick lava flows. In this case, the meltwater transport and fate processes will be essentially the same as with the initial very thick lava flow (Figure 5). This is because the melting interface will lie at the base of the lava sequence in contact with impermeable ice (Figure 7) (as was the case with the initial very thick lava flow), though less meltwater will be produced during the subsequent lava flow emplacement and heating (Figure 3). Throughout the top-down melting process the ice sheet will remain in a cold-based state (Fastook and Head, 2015) and the impermeable ice-cemented cryosphere will remain unaffected because the timescale of top-down melting (several 100 to 10^4 yr) is far more rapid than the timescale required for cryosphere reduction and melting (many 10^6 years). Due to the impermeable boundaries surrounding the lava flow sequence (Figure 7), meltwater is predicted to pool around the flow sequence, drain to the ice sheet base, or runoff across the ice sheet surface (Figure

7). However, since much less melt is being produced than during the initial thick lava flow emplacement, it is less likely that surface runoff will occur during subsequent lava flow emplacement. The fate of the water transported by these mechanisms remains the same as for the initial thick lava flow discussed in section 3.8. The top-down melting of the ice sheet, and evacuation of the meltwater, will result in substantial subsidence of the superposed lava flows which is predicted to lead to the formation of a host of subsidence and collapse related features including fracture systems, wrinkle ridges, depressions, and chaos terrain (Figure 7).

4.2 Subsequent Lava Flow Emplacement: Synthesis and Predictions

- Heat transfer from the emplacement of subsequent lava flows is delayed in delivery to the ice, reduced in magnitude, and more sustained in nature, relative to that from initial lava flow emplacement.
- Due to the reduction in heat transfer, the emplacement of subsequent lava flows will melt a smaller total thickness of ice relative to the initial lava flows.
- Meltwater produced by thinner subsequent lava flows (~10 m or less) is predicted to be predominantly absorbed into the surrounding firn layer.
- Top-down melting from subsequent emplacement of thinner lava flows is limited as successive flows are emplaced because heat delivery to the underlying ice is impeded by a thickening layer of previously emplaced chilled lava flows. Therefore, continued lava flow emplacement will result in negligible top-down melting, serving mainly to construct a thermally insulating cap of lava atop the ice sheet.

- Unless the ice sheets being loaded were initially very thick ($> \sim 500$ m), no ice will be left to melt from subsequent emplacement of very thick lava flows (~ 100 m or greater).
- If melting occurs from subsequent thick lava flow emplacement (~ 100 m or greater), meltwater is predicted to follow the same transport and fate processes outlined in section 3.7
- Melting of the underlying ice sheet by all thicknesses of subsequent lava flows will result in continued subsidence and degradation of the lava flow sequence (*e.g.* a 10 m lava flow, emplaced subsequent to an initial 10 m lava flow, will result in subsidence of the entire 20 m lava sequence on the order of ~ 15 m).
- At this point in the ice sheet lava loading process, lava flows will exist in a stable equilibrium condition on top of the ice sheet.

5. Continued Ice Sheet Lava Heating and Loading

While continued accumulation of thick lava flows will rapidly result in complete top-down melting of the predicted ice sheets, continued lava loading proceeds very differently if the accumulating lava flows are thin (~ 10 m or less). As thin lava flows continue to accumulate across an ice sheet surface, the portion of heat conducted down into the underlying ice is quickly reduced to the point that top-down melting at the ice sheet surface is negligible (If a 10 m lava flow is emplaced atop 90 m of cooled lavas, the underlying ice sheet will only undergo ~ 3 m of melting. If a 1 m lava flow were emplaced, only ~ 0.03 m of melting would take place). As a result of the limited top-down melting, the overall thickness of the ice sheet will not be significantly reduced, and the primary effect of the accumulating lava flows will be the establishment of a thermally

insulating layer atop the ice sheets, insulating the ice sheet and acting to raise the ice-melting isotherm. We now explore the long-term effects of continued ice sheet loading of relatively thin lava flows.

The distance that the ice-melting isotherm at the base of the cryosphere can be raised by the lava flows is dependent upon the thickness, and the thermal properties of the accumulated lava flows as well as the buried ice sheet. If a sufficient thickness of lava is loaded atop the ice sheets, the ice-melting isotherm may intercept the base of the ice sheet, resulting in induced ice sheet basal melting.

The lava heating and loading process and the nature of the induced melting processes are dependent on several critical factors: (1) The mean annual surface temperatures and the geothermal heat flux. (2) The thickness of both the cryosphere and the ice sheet upon which lavas are emplaced. (3) The total thickness to which lavas accumulate, and the timescale over which accumulation occurs. (4) The thickness of the individual lava flows, which has an effect on the amount of top-down melting produced during the lava heating and loading process. We review the effect of these factors in more detail in the following subsections.

5.1 Temperatures and Geothermal Heat Flux

Both the mean annual surface temperature and geothermal heat flux strongly influence the effect that the insulation provided by the lava loading process has on the cryosphere and the buried ice sheet. Higher temperatures and geothermal heat flux values allow for more rapid melting, while requiring less insulation, thus requiring lower thicknesses of accumulated lava. We test end-member cases for the mean annual surface temperature in the Hesperia Planum region during the Late Noachian of 210 and 240 K as

predicted by global climate modeling studies at atmospheric pressures of 8 mbar and 1 bar, respectively (Wordsworth et al., 2013). With respect to the geothermal heat flux, we assess a nominal Late Noachian geothermal heat flux of 55 mW/m^2 (Solomon et al., 2005; Clifford et al., 2010; Fastook et al., 2012), and an elevated geothermal heat flux of 100 mW/m^2 which may have been sustained, at least locally, during this time in the Hesperia Planum region due to widespread volcanic and magmatic activity (Cassanelli et al., 2015).

5.2 Ice Sheet and Cryosphere Thicknesses

Ice sheet thickness has a direct control on the bottom-up melting processes associated with the lava heating and loading mechanism in terms of the amount of ice available for melting, and in determining the thickness of loaded lava needed to cause bottom-up melting. This is because at greater ice sheet thicknesses more insulation is provided for the ice sheet base, and less additional insulation from accumulated lava is required to initiate bottom-up melting. The thickness of the cryosphere is determined by the balance between the mean annual surface temperature, the geothermal heat flux, and the thickness of the overlying ice sheet. Larger values of each of these parameters will reduce the thickness of the cryosphere.

The thickness of the regional ice sheets predicted to form across the highlands of Mars during the Late Noachian period (Wordsworth et al., 2013) depends on the total available surface water reservoir of Mars at this time (Carr and Head, 2015). It is likely that the Late Noachian available surface water reservoir was larger than the currently observed surface water inventory, currently $\sim 34 \text{ m}$ thick global equivalent layer (GEL) contained within the polar caps, and surface and shallow ground ice (Carr and Head,

2015). However, the precise quantity of water contained within the Late Noachian inventory depends on uncertain estimates of the amount of water lost to space (Greeley, 1987; Jakosky et al., 1994; Mellon and Jakosky, 1995; Greenwood et al., 2008), and to other sinks (*e.g.* to the deep groundwater system; Carr and Head, 2015) through time. To account for this, we assess a range of water inventory size scenarios. We test a surface water inventory equal to the current (~ 34 m GEL), and a water inventory 10X greater. Spread evenly across Mars above the predicted +1 km “icy highlands” equilibrium line altitude (Wordsworth et al., 2013), these surface water inventories produce average ice sheet thicknesses of 300 m and 1,000 m, respectively (Fastook and Head, 2015).

In order to model the cryosphere we adopt the Clifford (1993) crustal porosity (φ) structure which decays exponentially with depth (z) expressed as:

$$\varphi(z) = \varphi_o * \exp(-z/D) \quad (6)$$

where φ_o is the surface porosity, and D is a scaling factor for the decay of porosity with depth (estimated to be ~ 2.82 km for Mars; Clifford, 1993; Clifford et al., 2010).

Suggested martian surface porosities range generally from ~ 0.15 - 0.35 (Clifford, 1993; Hanna and Phillips, 2005; Clifford et al., 2010), with some estimates as high as 0.5 (Clifford, 1993). Here we adopt a relatively conservative surface porosity value of 0.2 .

While water-ice is stable within the frozen martian crust that comprises the cryosphere between the martian surface and the depth of the ice-melting isotherm, ice does not necessarily exist within the pore space of the subsurface. However, it is possible for ice to have accumulated within the pore-space of the Late Noachian cryosphere through diffusive transport of water from a deeper groundwater system (*e.g.* Clifford, 1991), or

for ice to have formed in situ by freezing of a saturated groundwater aquifer from an earlier “warm and wet” period (*e.g.* Craddock and Howard, 2002). In each model scenario tested, we assume the presence of an ice-cemented cryosphere (with ice completely occupying the available pore space), because it is likely for water to have been present at depth in the early Mars crust which would have contributed to, or been consumed in the establishment of a cryosphere by these processes. Under the variable temperature, geothermal heat flux, and initial ice sheet thickness conditions we explore, the ice-melting isotherm will lie between 0 to 2.5 km below the ice sheet base prior to lava heating and loading.

5.3 Lava Thicknesses and Accumulation Timescales

We adopt an individual lava flow thickness of 10 m, which is an average lava flow thickness observed in the Columbia River Flood basalt province (Self et al., 1997). We assess the range of total accumulated lava thicknesses (500 and 2,000 m) and lava accumulation timescales (10 kyr and 100 Myr) determined to be representative of the Hesperia Planum volcanic plains in section 2. To simulate lava accumulation, each emplacement timescale is divided into a number of time intervals equal to the ratio of the incremental lava flow thickness (10 m) divided by the total lava accumulation thickness (500-2,000 m), with a lava increment added at each time interval.

5.4 Lava Heating and Loading: Top-down Melting

We account for ice sheet top-down melting induced by supraglacial lava flow emplacement and cooling by implementing the parameterization derived in section 4 (equation 5) and setting the value of M_R to 3, corresponding to the 10 m thick lava flow case. Throughout the model simulation we assume that the top-down melting due to lava

flow emplacement occurs instantaneously. This assumption is made because in the early stages of lava heating and loading the top-down melting induced by incremental lava flow emplacement occurs over a time-scale that is approximately equal to the time step of the model. As subsequent flows accumulate, the induced melting begins to take place over a time greater than the model time-step, however at this point the melting is negligible, and the assumption of instantaneous melting is maintained.

5.5 Thermal Model

We assess the thermal evolution of the ice sheet, and subjacent cryosphere, in response to insulation from supraglacial lava heating and loading through the implementation of a fully explicit finite difference numerical scheme (*e.g.* Hu and Argyropoulos, 1996) to solve the one-dimensional heat conduction equation, expressed in terms of enthalpy as:

$$\frac{\partial H}{\partial t} = \frac{\partial}{\partial z} \left(k(z) \frac{\partial T}{\partial z} \right) \quad (7)$$

where, H is enthalpy (in J/kg), k is thermal conductivity (in W/m K), T is temperature (in K), and z is distance (in m). This equation is a reformulation of the standard heat conduction equation (*e.g.* Hu and Argyropoulos, 1996) which allows for modeling phase change problems by taking into account the latent heat of fusion. Each component of the system (lava, ice sheet, and cryosphere) is represented as a length within the one-dimensional model domain, the size of which corresponds to the thickness of the associated component. The lengths of the individual components are allowed to evolve with time in response to lava heating and loading and ice melting. A conceptual representation of the thermal model configuration is shown in Figure 8.

In each model run, the temperature at the upper model boundary ($z=0$) is held constant, while a constant geothermal heat flux is applied at the lower model boundary ($z=L$), the values of which are varied according to the model run scenario (Figure 8). The enthalpy is then calculated at each depth in the model domain using equation (7) and this is used to derive the local melt fraction based on a sharp melting front assumption (*e.g.* Alexiades and Solomon, 1992). The temperature at each point is then calculated from the associated enthalpy based on the phase state of the material as determined by the melt fraction (since the phase state of the material at any point in the model can change with time as a result of melting. *e.g.* consider a particular depth point in the model corresponding to the ice sheet, if the local melt fraction is zero, then no melting has yet occurred and the material is considered pure ice, and the enthalpy is converted to temperature accordingly). Throughout the modeling we maintain a spatial step size of 10 m (the largest spatial step that can be chosen while resolving the required features), and a time step of 0.5 years (the largest time step that can be chosen while maintaining numerical stability).

5.6 Thermal Properties

Over the range of temperatures involved in this analysis, the thermal conductivity of ice can vary appreciably. Therefore, we account for the temperature-dependent thermal conductivity of ice by the following (Alexiades and Solomon, 1992):

$$K_{T,ice}(T) = 2.24 + 5.975 \times 10^{-3} (273 - T(z))^{1.156} \quad (8)$$

where T is temperature (in K). In addition, we account for the temperature dependent thermal conductivity of the basaltic component of the porous substrate (Clifford et al., 2010) as:

$$K_{T,sub}(T) = 488.19/T + 0.4685 \quad (9)$$

where T is temperature (in K). For the lava flow material accumulating at the top of the ice sheet, we adopt an average of thermal conductivities measured from Hawaiian basalt flows and lunar basalt samples (~ 1.5 W/m K; Fujii and Osako, 1973; Robertson and Peck, 1974; Horai and Winkler, 1980; Warren and Rasmussen, 1987). We adopt typical density values of 917 kg/m^3 for ice and $3,000 \text{ kg/m}^3$ for rock/lava, as well as specific heat capacity values of $2,000 \text{ J/kg K}$ for ice and 850 J/kg K for rock/lava.

In the porous cryosphere, the thermal conductivity, specific heat capacity, and density are averaged by volume fraction between the pore ice and basalt substrate in accordance with the porosity (φ) such that:

$$K_{T,cry}(z) = \varphi(z) * K_{T,ice} + (1 - \varphi(z)) * K_{T,rock} \quad (10)$$

$$C_{P,cry}(z) = \varphi(z) * C_{P,ice} + (1 - \varphi(z)) * C_{P,rock} \quad (11)$$

$$\rho_{cry}(z) = \varphi(z) * \rho_{ice} + (1 - \varphi(z)) * \rho_{rock} \quad (12)$$

where the subscript *cry* denotes parameters associated with the cryosphere, and all other parameters are as previously defined.

5.7 Initial Conditions

Before any lava accumulation occurs, the temperature profile through the ice sheet and the substrate will be in a steady-state and nearly linear since the relatively thin ice

sheets (300 -1,000 m) will not flow rapidly (~12-431 mm/yr) under the cold Late Noachian conditions (Fastook and Head, 2015) of interest. The linear temperature gradient can be calculated using the steady-state solution of the one-dimensional heat equation:

$$\frac{dT}{dz} = \frac{z \cdot G}{\int_0^z K_T} \quad (13)$$

where:

z = depth (m)

G = geothermal heat flux (W/m²)

$\int_0^z K_T$ = integral of the thermal conductivity from the top of the lava flow to depth z (W/m K)

The temperature profile generated with this steady-state relation (subject to the same boundary conditions, and thermal conductivity parameterizations) is used to set the initial conditions for each model run.

6. Results

To explore the parameter space outlined in the previous subsections we perform a total of 32 individual thermal model runs. The array of model runs performed is shown schematically in Figure 9, with corresponding model run results compiled in Table 2.

In order to assess our findings, we define nominal cases from our range of modeled scenarios. We first take nominally predicted values for the mean annual surface temperature (210 K; Wordsworth et al., 2013), and geothermal heat flux (55 mW/m²; Solomon et al., 2005; Clifford et al., 2010; Fastook et al., 2012). With respect to the remaining parameters, two nominal scenarios are anticipated due to the predicted dichotomy in Hesperian volcanic plains emplacement conditions. In the Hesperia Planum

region, the topographically low craters would have acted as concentrating points for both ice (Fastook and Head, 2014; Fastook and Head, 2015) and lava (Whitten and Head, 2013). As a result, within a crater interior, the lava heating and loading process is predicted to be characterized by greater ice and lava thicknesses as well as more rapid lava accumulation rates. Therefore, in the nominal crater interior lava heating and loading scenario we assume an initial ice sheet thickness of 1 km, a total accumulated lava thickness of 2 km, and a lava accumulation timescale of 10 kyr. The final parameter required to define the nominal scenario is the thickness of the individual lava flows being accumulated, for which two possibilities exist. The lava flows accumulating in the crater may be relatively thin (~10 m thick) or thick (~100 m thick). The case in which the accumulating lava flows are relatively thick is described in section 4 in assessing the successive emplacement of very thick lava flows. Therefore, we now examine the case in which the individual accumulating lava flows are relatively thin (10 m thick). The thermal model output results for this defined nominal crater interior scenario (represented by model run 11; Figure 9) are displayed graphically in Figure 10.

Conversely, in the intercrater plains, ice thicknesses and accumulated lava thicknesses are predicted to be smaller, with lava accumulation taking place over a longer timescale. Therefore, in the nominal intercrater plain lava heating and loading scenario, we assume an initial ice thickness of 300 m, an accumulated lava thickness of 500 m, and a lava accumulation timescale of 100 Myr (represented by model run 5; Figure 9). The thermal model output results for the nominal intercrater plains scenario are displayed graphically in Figure 11.

In the nominal model run scenarios, as well as in all modeled scenarios, we find that as lava accumulates, the initial reduction in ice sheet thickness due to top-down melting defers the initiation of cryosphere bottom-up melting (*e.g.* Figure 10) (melting delay times are equivalent to the cryosphere melt initiation times listed in Table 2) by causing an initial decrease in the total insulation. However, as top-down melting of the ice sheet becomes negligible, the insulating effect of the accumulating lava flows becomes dominant, and the ice-melting isotherm begins to ascend towards the ice sheet base.

In the nominal crater interior scenario, we find that the additional insulation provided by the 2 km thick supraglacial lava flow sequence is sufficient to raise the ice-melting isotherm to the base of the superposed lava flow. As a result, the underlying ice sheet and cryosphere are rendered thermally unstable, and are subjected to melting as the ice-melting isotherm rises (Figure 10). In this case, the lava sequence is accumulated more rapidly than the ice-melting isotherm can rise, and thus the rate of bottom-up melting is limited by the geothermal heat flux input. Rapid lava accumulation and geothermally limited bottom-up melting can allow ice sheet basal melting to continue, and even begin, after lava accumulation has completed, we refer to this phenomenon as “deferred melting” (Figure 10). At the geothermally limited melting rates, melting of the initially ~1.9 km cryosphere takes place over a timescale of ~470 kyr, while melting of the ~830 m ice sheet (the amount that remains after top-down melting has completed) takes place over ~750 kyr.

In the nominal intercrater plains scenario, we find that the insulation provided by the 500 m thick supraglacial lava flow sequence is insufficient to raise the ice-melting isotherm to the base of the ice sheet, thus no ice sheet basal melting takes place. The

additional insulation is only sufficient to raise the ice-melting isotherm from an initial depth of 2.5 km, to a depth of 1.8 km resulting in cryospheric melting over the same depth range (Figure 11). In this case, melting takes place over a timescale of ~ 82 Myr because the rate at which the ice-melting isotherm can rise is limited by the rate at which insulation is provided from lava accumulation (which now occurs over a 100 Myr timescale). This is because after each increment of lava is emplaced, the resultant elevation of the ice-melting isotherm and associated ice melting occurs before the next increment of lava is added. As a result, bottom-up melting occurs in steps (Figure 11), with short periods of more rapid geothermally-limited melting just after an increment of lava is added, followed by a period of no melting until more insulation is provided by the emplacement of the next lava flow. As a result of this limitation, over the course of lava accumulation (100 Myr), the long-term averaged bottom-up melting rate of the cryosphere is equal to the lava accumulation rate.

The effects of variable surface temperatures, geothermal heat flux, and lava heating and loading conditions are explored in the remaining model run scenarios (Figure 9; Table 2). In general, we find that at the predicted intercrater plain ice sheet thickness of 300 m, basal melting is only possible with the addition of the 2 km thick lava flow sequence, except in cases with both a very high surface temperature (240 K, predicted to occur at an atmospheric pressure of 1 bar; Wordsworth et al., 2013) and geothermal heat flux (100 mW/m^2 , reflective of a regionally enhanced geothermal heat flux; Cassanelli et al., 2015). Conversely, at an ice sheet thickness of 1,000 m, basal melting is predicted in all lava heating and loading scenarios except those with a lower surface temperature (210 K), lower geothermal heat flux (55 mW/m^2), and thin 500 m layer of superposed lava

(Table 2). This is in contrast to the ice-loading alone case (Cassanelli et al., 2015) in which no basal melting is predicted to occur, even with plausible regionally elevated geothermal heat fluxes.

We find that geothermally-limited rates of bottom-up melting-front advance in the cryosphere and ice sheet are on the order of ~ 5 mm/yr and ~ 2 mm/yr, respectively. When accounting for porosity and density effects, these melting front advance rates translate into long-term bottom-up meltwater production rates of ~ 0.5 mm/yr in the cryosphere and ~ 1.8 mm/yr in the ice sheet (typical basal melting rates for terrestrial glaciers are on the order of several millimeters per year, though melting can be significantly enhanced in volcanically active regions, and has been reported to be as high as ~ 5 m/yr in some portions of the Vatnajökull ice cap in Iceland; Cuffey and Paterson, 2010). Long-term lava accumulation-limited rates of melting front advance in the cryosphere and ice sheet are both on the order of ~ 0.02 mm/yr (since both are limited by the rate at which insulation is provided by lava accumulation). These long-term averaged melting front advance rates translate to long-term averaged meltwater production rates of ~ 0.002 mm/yr in the cryosphere, and ~ 0.018 mm/yr in the ice sheet.

The maximum bottom-up melting front advance rates predicted in this study are ~ 15 mm/yr in the cryosphere (achieved in model run 20; Figure 9) and ~ 6 mm/yr in the ice sheet (achieved in model run 28; Figure 9). These rates are achieved in scenarios where a large increase in ice sheet insulation is provided by rapid accumulation (10 kyr) of 2 km of lava atop the ice sheet, with subsequent bottom-up melting proceeding at a rate governed by the highest geothermal heat flux evaluated in this study (100 mW/m^2).

6.1 Bottom-up Induced Melting: Cryosphere Contribution

In the nominal crater interior lava heating and loading scenario we explore (Figure 10), the cryosphere underlying the ice sheet initially extended to a depth of ~ 1.9 km. After the insulating lavas are loaded atop the surficial ice sheet, the equilibrium thermal state is interrupted, and the ice-melting isotherm which defines the extent of the cryosphere, advances towards the surface. If the cryosphere were ice-cemented this would result in melting and the liberation of meltwater to the underlying substrate. In this scenario, the entire 1.9 km thick cryosphere is completely removed due to the large amount of insulation provided. Under the assumed porosity structure (section 5.2), and assuming the cryosphere was initially ice-cemented, melting of the 1.9 km thick cryosphere would result in the release of ~ 250 m column of meltwater per unit area. Within the Hesperia Planum region, the largest observable craters are on the order of ~ 80 km in diameter (Tanaka et al., 2014b), melting from this process, in a crater of this scale, would release ~ 0.03 m GEL of water to the subsurface in each crater of this size.

In the nominal intercrater plains scenario (Figure 11), the cryosphere underlying the ice sheet initially extended to a depth of ~ 2.5 km due to the reduced insulation from the thinner surficial ice sheet. Following the lava heating and loading predicted in this scenario, the cryosphere is thinned to a depth of ~ 1.8 km. Under the assumed porosity structure (section 5.2), and assuming the cryosphere was initially ice-cemented, this would result in the liberation of ~ 60 m column of meltwater per unit area. If this melting occurred over the entire Hesperia Planum region, a ~ 1 m GEL of water would be released to the subsurface.

Under the full range of conditions explored here (Figure 9; Table 2), prior to any ice sheet lava heating and loading, the cryosphere will extend from ~ 0 -2.5 km below the

surface. We find that model runs 3 and 7 produced the greatest extent of cryosphere melting (Table 2). In these scenarios, the cryosphere was initially 2.5 km thick (due to the low mean surface temperature, low geothermal heat flux, and thin ice sheet). Subsequent accumulation of a thick sequence of lava flows (2 km), resulted in complete bottom-up melting of the cryosphere. Given the assumed crustal porosity structure, if the cryosphere were initially ice-saturated this would release a ~331 m column of meltwater per unit area. If this melting occurred over the entire Hesperia Planum region, a ~4.5 m GEL of water would be released to the substrate. In the scenarios with a high mean annual surface temperature, high geothermal heat flux, and thick ice sheets (e.g. model run 26), the ice-melting isotherm is predicted to lie at the base of the ice sheet such that no cryosphere is predicted to exist. Therefore, in these model runs, no meltwater is generated through cryospheric melting.

Given that cryospheric ice melting would occur at depth within the martian subsurface, any meltwater produced by bottom-up melting of an ice-cemented cryosphere would drain further into the subsurface (if the underlying material is permeable) (Figure 12). Thus, the fate of this meltwater will be dependent upon the presence, or absence, of a more extensive groundwater system deeper within the martian crust. If a groundwater system does exist at depth, then it must be at diffusive equilibrium with the ice-cemented cryosphere above, such that the cryosphere is ice-cemented to the depth of the ice-melting isotherm (*e.g.* Clifford, 1991; Mellon et al., 1997). If this was not the case, then any groundwater present would have undergone vapor diffusion (Clifford, 1991) and have become sequestered within the ice-cemented cryosphere (thereby thickening the ice-cemented cryosphere and bringing the ice saturation line closer to the depth of the ice-

melting isotherm). Therefore, if a groundwater system is present at greater depth, then the meltwater from melting of the ice-cemented cryosphere will simply provide groundwater recharge for that system and enter storage within the aquifers. Alternatively, if an extensive groundwater system does not exist, then this water would move down until the subsurface became impermeable at which point it would begin to migrate laterally, initiating aquifer formation. Once enough water infiltrates, the aquifer will spread beyond the bounds of the surficial lava flows where the cryosphere will again be stable to great depth. Here, the cryosphere may not be ice-cemented to the depth of the ice-melting isotherm, in which case water from the newly formed aquifer system would undergo diffusive loss until either the groundwater is depleted or the ice cementation line reaches the ice-melting isotherm depth, establishing vapor diffusive equilibrium (Clifford, 1991).

6.2 Bottom-up Melting: Ice Sheet Meltwater Transport and Fate

In the two nominal scenarios we explore (crater interior and intercrater plains lava heating and loading; Figures 10 and 11), bottom-up ice sheet melting as a result of lava heating and loading is predicted to occur predominantly within crater interiors (since ice sheet and lava thicknesses we adopt in the intercrater plains are too low to provide sufficient insulation for basalt melting). In the nominal crater interior scenario, the thick (2 km) accumulation of lava causes the ice-melting isotherm to ascend to the base of the lava flows, leading to complete melting of the underlying ice sheet (of the 1 km thick ice sheet 170 m is removed through initial top-down melting, with the remaining 830 m melt through bottom-up basal melting). As a result, a ~760 m column of melt water is produced per unit area of melting which would produce a ~0.1 m GEL of water within an 80 km diameter crater. While lava heating and loading induced basal melting is not

predicted in the nominal intercrater plains scenario, it is possible for basal melting to have taken place if a greater thickness of lava (2 km) were accumulated atop the ice sheets, or if the surface temperature and geothermal heat flux were considerably higher (240 K, and 100 mW/m²). If basal melting in the intercrater plains occurred as a result of these conditions, top-down melting would remove ~130-170 m of the ice sheets, leaving also 130-170 m to undergo basal melting. Complete basal melting of these ice sheet thicknesses would release ~120-155 m column of meltwater per unit area which over the area of Hesperia Planum would release a ~1.6-2.1 m GEL of meltwater.

Broadly, there are two possible fates for meltwater released by ice sheet basal melting: (1) meltwater may infiltrate down into the porous substrate (Figure 12), or (2) meltwater may become sequestered beneath the ice and lava flows due to the presence of an impermeable underlying layer (Figure 12). Since basal melting initiated by the lava heating and loading process occurs through bottom-up heating, the cryosphere is predicted to complete melting before ice sheet basal melting can occur since the ice-melting isotherm cannot rise beyond the melting front. Therefore, the meltwater produced by ice sheet melting will not encounter an impermeable ice-cemented cryosphere, although it is possible for other impermeable layers to exist beneath the ice sheet (*e.g.* a clay layer or competent bedrock; Figure 12).

The rate at which the meltwater generated at the base of the ice sheet can infiltrate into the subsurface is governed by the infiltration capacity of the substrate material. To estimate the infiltration rate, we apply the adaptation of Darcy's law for saturated groundwater flow described by equation (2) in section 2.1 subject to the same assumptions. These assumptions again reduce equation (2) to give the saturated hydraulic

conductivity as the infiltration rate. As noted before, the saturated hydraulic conductivity will be an overestimate if the substrate beneath the ice sheet is in a desiccated state. However, in this case, melting of an ice-cemented cryosphere prior to ice sheet basal melting would result in conditions closer to the saturation point, thus the hydraulic conductivity may be closer to, or even at the saturated value. To estimate the hydraulic conductivity of the martian substrate we adopt the Clifford and Parker (2001) permeability structure and apply the relationship for intrinsic permeability and hydraulic conductivity detailed in section 2.1. From this process, we find hydraulic conductivity values of ~ 75 mm/hr at the surface, decreasing to ~ 0.007 mm/hr at 2 km depth after which point the hydraulic conductivity remains nearly constant to the self-compaction depth at ~ 10 km (Hanna and Phillips, 2005). As the substrate saturates, the infiltration will become limited by the hydraulic conductivities of the material at greater depth, such that the conductivity at the surface is not representative of the actual infiltration rate. For comparison the infiltration rates of terrestrial volcanic terrains with a similar permeability structure has been measured at ~ 0.1 mm/hr (Hurwitz et al., 2003). However, since the infiltration rate is proportional to gravity, this infiltration rate should be closer to $0.1 \text{ mm/hr} * (g_{\text{Mars}}/g_{\text{Earth}})$ or ~ 0.03 mm/hr on Mars.

These infiltration rates are more than sufficient to accommodate the maximum ice sheet basal melting rates predicted in the most extreme melting scenario modeled (model run 28), which are $\sim 7 \times 10^{-4}$ mm/hr. Only if infiltration becomes limited to the lowest permeability ($\sim 10^{-15} \text{ m}^2$) material at great depth in the martian mega-regolith (which is not predicted to occur since the exceedingly low melting rates will not be able to saturate the substrate to these depths) will the saturated hydraulic conductivity ($\sim 7 \times 10^{-6}$ mm/hr)

be below the highest predicted melting rates ($\sim 7 \times 10^{-4}$ mm/hr). Therefore, we predict that the meltwater produced by ice sheet basal melting during the lava heating and loading process will percolate into the martian substrate providing groundwater recharge (Figure 12).

Meltwater will continue to percolate downwards until it joins a deeper groundwater system or encounters an impermeable layer at which point it will begin to establish an aquifer. If the available pore space in the substrate does not extend laterally beyond the bounds of the melting zone, it is possible for the meltwater to exceed the storage capacity of the substrate. Given our assumed porosity structure, we estimate that a ~ 10 km column (to the self-compaction depth; Hanna and Phillips, 2005) of substrate material can accommodate meltwater produced from a ~ 600 m column of ice. Therefore, unless the meltwater is able join an aquifer system whose bounds extend laterally beyond the zone of melting, the storage capacity of the aquifer would be exceeded prior to complete ice sheet melting in the nominal crater interior scenario. As a result, additional meltwater would then pool at the base of the melting ice sheet. This will be compounded if the cryosphere were ice-cemented to the depth of the ice-melting isotherm (~ 1.9 km), which would reduce the storage capacity to accommodate a ~ 320 m column of ice. If the groundwater at depth beneath the global cryosphere is able to flow out past the ice sheet margins then additional melt can be accommodated by the aquifer. Even if the meltwater is able to flow out to a more extensive groundwater system beyond the zone of melting, it must do so at depth beneath the cryosphere where the permeability is relatively low ($\sim 10^{-14}$ to 10^{-16} m²). At these permeabilities the hydraulic conductivity can be as low as 0.0065 m/yr which is still sufficient to accommodate the basal melting rates predicted in the

nominal crater interior scenario ($\sim 9 \times 10^{-4}$ m/yr), as well as the maximum melting rates predicted in the most extreme bottom-up melting scenario assessed (~ 0.006 m/yr). Therefore, the pooling of meltwater beneath the lava-loaded “icy highlands” is not predicted given the nominal subsurface permeability structure.

While the ice sheet basal melting rates are not predicted to exceed the infiltration capacities associated with the nominal permeability structure, the presence of low permeability, or impermeable layer within the substrate could impede meltwater infiltration (Figure 12). If this is the case, meltwater will pool beneath the ice sheet and will not be able to escape because the margins of the ice sheet will remain frozen to the base. Water that pools beneath the ice sheet will be under a large confining pressure due to the overburden stress of the ice and lava flows above. If a sufficient pressure is built up the ice sheet may fracture at the margins where the ice is thin, resulting in a flooding event (Figure 12). A flood event produced in this way would release a substantial quantity of water at very significant pressures. Considering the nominal crater interior lava heating and loading scenario, a 10 m meltwater lens within an 80 km crater would contain $\sim 200 \text{ km}^3$ of water, and beneath 2 km of lava and ~ 820 m of ice, would be under pressures of ~ 25 MPa (which is on the order of the tensile, $\sim 0.7\text{--}3.1$ MPa, and compressive, $\sim 5\text{--}25$ MPa, strength of ice under similar temperature regimes; Petrovic, 2003). A flood event produced through this mechanism would form outflow channels near the ice sheet margin (Figure 12) and would leave a large void space at the ice sheet base which could cause collapse of the superposed ice and lavas leading to the formation of collapse features (*e.g.* Zegers et al., 2010).

Regardless of whether the meltwater produced from ice sheet basal melting is evacuated by episodic flooding, or by gradual subsurface infiltration, the superposed lava flows will undergo subsidence as a result (Figure 12). However, the subsidence of the superposed lava flows will be affected by the timescale of lava accumulation and melting. For example, in the nominal crater interior scenario, the entire lava sequence is predicted to accumulate prior to the initiation of bottom-up ice sheet melting. As a result, the entire sequence of flows will undergo subsidence of ~830 m, as the ice sheet remaining after top-down melting is removed. Conversely, if lavas are added slowly, bottom-up melting will occur in increments following the addition of each lava increment, and completing before the addition of the next lava increment. Therefore, subsidence will occur in the same incremental nature, such that different portions of the total lava sequence will be processed by different amounts of subsidence. The incremental nature of subsidence means that flows emplaced early in the lava loading process will undergo more total subsidence than flows emplaced at the end of the process, when little ice is left to melt. Therefore, if lava loading occurred slowly, the remaining lava flows observed at the surface will not have been highly processed by subsidence. However, if lava loading occurred rapidly and completed before bottom-up ice sheet melting could take place, then the surficial lava flows would be highly processed as a result of large subsidence.

Subsurface mass loss from the buried ice sheet, and the associated subsidence of the superposed lava flows could lead to the development of a range of associated subsidence and collapse features expressed at the surface of the superposed lava sequence (Figure 12). These features include chaos terrain (*e.g.* Zegers et al., 2010), fracture systems, pit crater chains (*e.g.* Wyrick et al., 2004), and linear and irregularly shaped folding,

buckling, and faulting of the lava flow surfaces (*e.g.* wrinkle ridges, arches, normal faults, deformation rings). Potential examples of these features, found within the Hesperia Planum region, are documented in Figure 13.

After bottom-up melting has completed, the cryosphere will be reestablished through vapor diffusion. In the nominal cases explored, the reestablished cryosphere will extend ~1.7-2.3 km below the surface. Since the fractured lava flows from the lava loading process now occupy the upper 500-2,000 m of the surface (Figure 7 & 12), an equivalent thickness of the cryosphere will now be reestablished within this material which will have effectively formed a fractured rock aquifer. This fractured rock material will replace the mega-regolith material which initially contained the cryosphere causing a net decrease in the amount of water required to re-establish the cryosphere since fractured rock aquifers generally contain less available pore space than comparable mega-regolith aquifers (Hanna and Phillips, 2005).

6.3 Pressure Melting Point Reduction

We have assumed throughout this assessment that the melting temperature of ice remains constant at 273 K. However, the overburden pressures generated by the ice sheet and the accumulated lava can reduce the melting point of the ice at the base of the ice sheet to as low as ~271 K (at a pressure of ~25 MPa in the case where 2 km of lava are superposed upon 1 km of ice). Limitations in the numerical model prevent direct consideration of this effect, therefore we briefly discuss the implications of pressure melting point reduction. While the overburden pressure produced by the weight of the lava sequence and ice sheet reduces the melting temperature at the base of the ice sheet, this effect is not predicted to propagate to the ice contained in the pores of the substrate,

because the substrate matrix supports the overburden load. As a result, a discontinuity in melting temperatures will exist at the interface between the ice sheet and the substrate, with the ice at the base of the ice sheet requiring a lower temperature to initiate melting. This allows the ice sheet to begin bottom-up basal melting, before melting of the underlying cryospheric pore ice has completed. As a result the substrate would remain impermeable during the initial stages of bottom-up ice sheet melting, causing meltwater produced from ice sheet basal melting to pool at the base of the ice sheet, forming a melt lens. Given the geothermal gradients we test here (55 mW/m^2 and 100 mW/m^2), this effect can allow a melt lens ranging from $\sim 45\text{-}80 \text{ m}$ thick to form prior to complete cryosphere removal. The development of a melt lens at the base of the ice sheet would result in the initiation of wet-based glaciation, leading to enhanced ice sheet flow and basal erosion. In addition, flooding events could be triggered by the release of water contained in the melt lens, potentially resulting in large-scale deformation and collapse of the superposed lavas due to the rapid excavation of underlying material and the creation of a subsurface cavity.

6.4 Effect of Ice Impurities

In the analyses performed here, we make the assumption that the ice involved is free of impurities. In reality it is probable that the pore ice in the cryosphere, and the ice comprising the ice sheets, would have contained some component of impurities. These are likely to include volcanic ash, dust, and potent freezing point depressing salts (such as perchlorates, sodium chloride, and calcium chloride; Clifford et al., 2010). The precise composition and concentrations of impurities within the ice are unclear and due to this uncertainty, we do not directly account for impurities within the numerical model applied

here. While the incorporation of impurities within the ice is not predicted to significantly alter the results of the lava heating and loading process, there will be minor effects with respect to the associated top-down and bottom-up melting processes which we now discuss.

Top-down Melting: Incorporation of impurities into the ice sheet will have two main effects on top-down melting associated with supraglacial lava flow emplacement. The impurities will (1) reduce the melting temperature of ice, allowing for more melting, and (2) will diminish the thickness of the ice sheet firn layer (Cassanelli and Head, 2015). As a result, firn absorption may no longer be a major pathway available for meltwater produced from conductive heating during supraglacial lava flow emplacement and cooling. Instead meltwater will primarily be collected/transported by: (1) Drainage down through the ice sheet (e.g. through fractures). (2) Runoff across the ice sheet surface, following ice sheet topography. This will result in channelization across the ice sheet surface, and possibly in the formation of fluvial channels at the ice sheet margin when the melt drains off the ice sheet. (3) Pool around the emplaced lava flow, enhancing cooling rates and the likelihood of pseudo-crater formation. Regardless of these effects, top-down melting will still become rapidly diminished as lava accumulation continues, and the end result of the process will still be the construction of a thermal blanket atop the ice sheet.

Bottom-up Melting: The inclusion of impurities into cryospheric ice and into the ice sheet will serve to enhance the effect of bottom-up melting induced by the lava heating and loading process by depressing the melting point of the ice. This reduction in the melting temperature will allow for melting to initiate at a lower thickness of insulating lava flows, and will enhance the rate of melting front advance. If impurity concentrations

are higher within the ice sheet than in the cryospheric ice, it may also add to the pressure melting point reduction effect, enhancing the ability of the ice sheet to undergo basal melting prior to complete cryospheric melting. This would lengthen the window of time during which an entirely impermeable ice-cemented substrate would exist beneath the melting ice sheet, causing more pooling of meltwater at the ice sheet base.

6.5 Effect of Coeval Ice and Lava Accumulation

Throughout this assessment it has been assumed that ice sheet formation was completed prior to lava flow emplacement. However, ice and lava may have instead accumulated synchronously, resulting in interspersed deposits of ice and lava. This may even have occurred as a direct consequence of the ice sheet lava heating and loading process. If melt liberated from the highlands ice sheets was transported to the lowlands it could have been recycled back to the highlands by evaporation and re-deposition as described in the “icy highlands” climate scenario prior to the cessation of lava emplacement. Interspersed deposition of ice and lava resulting from coeval accumulation would influence the heat delivery and melting mechanisms explored in the ice sheet lava heating and loading model and would be dependent upon the balance between the rate of ice and lava accumulation. Given that ice deposition is predicted to occur at an average rate of ~ 10 mm/year under “icy highlands” climate conditions (Forget et al., 2013; Wordsworth et al., 2013; Fastook and Head, 2015), the influence of coeval accumulation would be governed by the rate of lava accumulation.

In general, three cases are possible. (1) Lava flows undergo rapid accumulation over a period of ~ 10 kyr (as discussed in Section 2). In this scenario, the introduction of snow at a rate of ~ 10 mm/year would only allow ice deposits on the order of ~ 1 m thick to

accumulate in between lava flow emplacement events. As a result, the deposited snow and ice would not accumulate to a substantial thickness and would undergo melting and evaporation without having a significant effect on the cooling processes of the lava flows. Therefore, the processes and morphology predicted by the ice sheet lava heating and loading model would not occur, although phreatomagmatic interactions would be possible. (2) Lava flows undergo gradual accumulation over a period of ~100 Myr (as discussed in Section 2). In this case, much larger periods of time would exist in between lava flow emplacement events, allowing snow and ice to accumulate to greater thicknesses. The average time in between lava flow emplacement events in the gradual accumulation scenario could be as much as ~2 Myr (in the intercrater plains where 500 m of lava is accumulated from 10 m lava flows over 100 Myr). In this period of time, snow deposition at an average rate of 10 mm/year would readily deplete even the maximum plausible Late Noachian/Hesperian surface water reservoir (~10 X the present, or ~340 m GEL; Carr and Head, 2015) resulting in complete supply-limited ice sheet formation. Therefore, subsequently emplaced lava flows would encounter a fully-formed ice sheet as is treated in our analyses, resulting in the previously described ice sheet lava heating and loading processes and morphology. (3) Accumulation rates of lava and snow/ice are comparable in magnitude. In this case each emplaced lava flow would encounter a layer of snow or ice that is on the same order of thickness as the lava flow itself. Under these conditions each lava flow would be an initial lava flow, in accordance with our previous definition. Therefore, the interaction of each sequentially emplaced lava flow with the snow and ice deposits would proceed as described in Section 3, resulting in the generation of the predicted morphology.

We find that the processes of heat delivery and melting involved in a scenario of coeval accumulation of snow/ice and lava are effectively described by the processes and conditions we have treated. However, interspersed deposition of snow/ice and lava could lead to an enhancement in aqueous/thermal mineralogic alteration of the basaltic lavas relative to the nominal case (in which ice sheet formation predates lava flow emplacement) due to more direct interaction between each freshly emplaced lava flow with ice and water.

6.6 Lava Heating and Loading: Synthesis and Predictions

- Due to the highly cratered nature of the topography onto which the Hesperian Planum volcanic plains were emplaced, a dichotomy in lava emplacement conditions is predicted between crater interiors and intercrater plains.

Nominal Crater Interior Lava Heating and Loading:

- Crater interiors are predicted to act as concentrating locations for both ice and lava. As a result the lava heating and loading process within crater interiors will be characterized by greater thicknesses of ice and lava, and by more rapid accumulation of lava.
- The greater thickness of lava accumulated within crater interiors provides sufficient insulation to lift the ice-melting isotherm to the base of the lava flows superposed on the ice sheet. This results in complete thermal instability of the cryosphere and ice sheet, and eventual melting governed by the geothermal heat flux. This process can produce ~0.1 m GEL of water within an 80 km crater.

- Due to the rapid nature of lava accumulation within crater interiors, bottom-up ice sheet melting can continue, or even begin, after lava accumulation has completed, leading to “deferred melting” occurring as much as 871 kyr years later.
- The maximum predicted bottom-up melting rates are far below the infiltration rates calculated for the martian substrate, and thus meltwater is predicted to percolate into the subsurface.
- Subsurface mass loss from bottom-up ice sheet melting will cause the superposed lava flows to undergo a great deal of subsidence on the order of several hundred meters. This is predicted to cause extensive fracturing of the superposed lavas and to give rise to a host of associated deformation and collapse features including chaos terrain, fracture systems, wrinkle ridges, pit crater chains, and depressions.
- If an impermeable layer of material (e.g. clay or competent bedrock) underlies the area of bottom-up melting, meltwater may become sequestered at the ice sheet base. Melt sequestered at the ice sheet base could be confined by very large overburden pressures from the superposed lava sequence. This water could be released through ice sheet fracturing near the ice sheet margins resulting in large-volume episodic flooding events.
- A flood event produced through this mechanism could form outflow channels which would emerge from the lava plains near the ice sheet margin. In addition this process could form a large void space at the ice sheet base which would cause collapse of the superposed ice and lavas leading to surface deformation and the formation of collapse features in the source region of the meltwater.

- Pressure melting point reduction, and the inclusion of impurities, could allow the ice sheet to undergo basal melting prior to complete melting of the underlying cryosphere. In this case the cryosphere would then act as an impermeable barrier, preventing the downward infiltration of melt from the ice sheet, having the same effect as other impermeable formations.

Nominal Intercrater Plains Lava Heating and Loading:

- Lava heating and loading in the intercrater plains is nominally predicted to be characterized by thinner accumulations of ice and lava, and by more gradual lava accumulation.
- Due to the reduced thicknesses of ice and lava predicted, insufficient insulation is provided in the intercrater plains to raise the ice-melting isotherm to the base of the ice sheets. As a result, only limited cryosphere melting is predicted.
- Given that none or minimal ice sheet basal melting or subsurface mass loss is predicted in the nominal intercrater plains, minimal subsidence or collapse related features are predicted to form.
- Basal melting in the intercrater plains is possible, even at the thin predicted ice sheet thicknesses, if greater thicknesses of lava were accumulated (2 km), or if the mean annual surface temperature and geothermal heat flux were considerably higher (240 K, and 100 mW/m², respectively).
- Subsurface mass loss from buried ice sheet basal melting in the intercrater plains will result in the formation of the same suite of subsidence and collapse features as predicted for the crater interior scenario.

7. Example Case

In order to test a physical application of the lava heating and loading concept, we apply the lava heating and loading model to a small study area located in the northern portion of the Hesperia Planum region centered approximately at 106.496 °E and 6.205 °S (Figure 14). Within this region lie two impact craters, ~27 km in diameter (Crater A; Figure 14), and ~22 km in diameter (Crater B; Figure 14), which are flooded and mapped as part of the Hesperian Ridged plains unit (Tanaka et al., 2014b). The filling unit within each impact structure shows evidence for post-depositional modification in the form of wrinkle ridges and fracture systems (Figure 14). A broad (~3 km wide) channel emerges from the rim along the northwestern portion of the northwestern crater (crater A; Figure 14) which contains tear-drop shaped islands. Due to the range of evidence suggesting post-depositional modification of the Hesperian Ridged plains filling unit within the two impact craters contained in the study region, we assess this area in the framework of the ice sheet lava heating and loading model to determine if the model can explain the presence of the observed features.

To perform this assessment we first assume that the filling unit of each impact crater is comprised entirely of Hesperian Ridged plains volcanic material. Maximum accumulated lava thicknesses are then calculated by the same manner described in section 2, yielding maximum lava thicknesses of ~1.5-1.75 km. We continue by assuming that these impact crater structures would have acted to concentrate both ice and lava deposits based on the same justifications provided in section 6. We therefore assume that a 1 km thick ice deposit existed in the crater before lava emplacement, and that lava emplacement occurred over a rapid timescale (10 kyr). With these assumptions, we now outline two possible lava heating and loading scenarios for the study area. (1) In the first

scenario, lava accumulation is assumed to have occurred through the emplacement of a few very thick lava flows (~100 m thick or greater) (Figure 7). (2) In the second scenario, lava accumulation is assumed to have occurred through the emplacement of a greater number of thinner lava flows (~10 m) (Figure 12).

In scenario (1), complete melting of the 1 km thick ice sheets is predicted to occur by top-down melting through conductive heat transfer from the emplaced lava flows (Figure 3). If the lava fill were accumulated from individual 100 m thick lava flows, then complete melting of the ice sheet would occur within the emplacement of ~5 flows, while if the individual lava flows were 200 m thick, one flow is predicted to transfer enough heat to melt nearly the entire 1 km thick ice sheet (Figure 3). In either case, complete melting of the ice sheet would occur over rapid time scales, on the order of several 100 to 10^4 years. Due to the rapid nature of heat transfer and melting in this scenario, there will not be sufficient time to allow for thinning and removal of the underlying ~1.9 km thick cryosphere. As a result, the subsurface will remain impermeable throughout the lava heating and loading and melting processes (Figure 7). Since the meltwater cannot percolate into the subsurface it is predicted to pool at the base and margins of the emplaced flow, which could overtop or rupture the confining ice leading to large episodic meltwater flooding events at the volcanically flooded crater margins (Figure 7). Melting and removal of the buried ice sheet would result in subsidence of the superposed lava flows on the order of ~1 km gradually over the course of the melting period (several 100 to 10^4 years). However, more rapid subsidence events could result from episodic release of confined meltwater, potentially generating deformation in the source region of the water. Therefore in this scenario the generation of several characteristic features are

predicted to result from the lava heating and loading process (Figure 7 & 13): (1) fractures and cracks within the superposed lavas due to subsidence, (2) collapse features including depressions, chaos terrain, or pit craters associated with subsurface mass loss, and (3) fluvial channels emerging from the lava flow margins associated with episodic meltwater release in flooding events.

Scenario (2) is effectively represented by the nominal crater interior lava heating and loading scenario outlined in section 6. In this case, top-down melting is limited due to the inefficient transfer of heat resulting from successive emplacement of the relatively thin 10 m thick lava flows. As a result, only ~170 m of total top-down ice sheet melting occurs during the emplacement of the 2 km thick lava sequence. The insulation provided by the 2 km thick lavas is predicted to lift the ice-melting isotherm to the base of the lava flows. Therefore, following lava emplacement and cooling, the initially 1.9 km thick cryosphere underlying the ice sheets will undergo melting over ~0.5 Myr, followed by complete “deferred melting” of the remaining ~830 m buried ice sheets over a time scale of ~0.75 Myr (Figure 12). The maximum bottom-up melting rates predicted (~1 mm/yr) in this lava heating and loading scenario fall far below the predicted infiltration capacity of the nominal substrate (although the permeability of the crater floor materials relative to the nominal mega-regolith is unclear because heavy fracturing associated with the impact process would increase the predicted permeability and infiltration rates, but the presence of a solidified impact melt sheet could serve to reduce them). Therefore, in this scenario we predict rapid accumulation of lavas resulting in a limited amount of concurrent top-down melting, followed by complete bottom-up melting of the cryosphere and ice sheet limited by the rate at which heat is delivered from the geothermal gradient (Figure 10).

Meltwater resulting from bottom-up melting in this scenario is predicted to infiltrate into the subsurface to provide groundwater recharge (Figure 12). However, the overburden pressure from the loaded lavas may have initially allowed ice sheet basal melting to begin prior to cryospheric melting due to the pressure melting point reduction effect discussed in section 6.4. This could have allowed for the accumulation of a melt lens at the ice sheet base which if released through rupturing of the confining material, would have caused a flooding event. However, a flooding event produced in this way would occur only during the onset of bottom-up ice sheet melting, since after the flooding event, the ice-melting isotherm would have ascended into the ice sheet, thus removing the impermeable cryosphere. Features predicted to result from this second lava heating and loading scenario include (Figure 12 & 13): (1) Primarily fractures and cracks within the superposed lavas due to gradual subsidence from long term subsurface ice mass loss and meltwater infiltration. (2) Collapse features including depressions, chaos terrain, or pit craters are possible due to early events of more rapid subsurface mass loss from flooding events facilitated by pressure melting point reduction effect. However, these features could only form during a brief window of time at the onset of ice sheet basal melting. Therefore, these features are not predicted to be dominant, and would additionally show evidence for further subsidence resulting from subsequent gradual subsurface mass loss from ice sheet basal melting and melt infiltration.

(3) Fluvial channels emerging from the lava flow margins associated with flooding events. Fluvial channels in this scenario would be predicted to be generally small in scale due to the limited nature of the conditions which could produce flooding.

Examination of the features observed within the study region (Figure 14) indicates the presence of: (1) large cracks and fracture systems within the volcanic crater fill units (Figure 14), (2) a large (~3 km wide) channel emanating from the rim of the Northwestern crater (crater A), which we interpret to be fluvial in origin as suggested by the presence of tear-drop shaped islands within the main channel (Figure 14). The cracks, fracture systems, and wrinkle ridges observed within the volcanic filling unit in each crater are generally consistent with either lava heating and loading scenario and in this example do not allow clear distinction between the models. However, the large scale of the channel observed to emerge from the rim of the crater is suggestive of more rapid and significant discharge of water which is predicted to result predominantly from the conditions outlined in the first scenario in which the emplacement of very thick lava flows (~100 m or greater) result in significant and rapid top-down melting. The observed geological evidence (Figure 14) is generally consistent with the predictions made by lava heating and loading scenario (1) (Figure 7). Therefore we predict that lava heating and loading of the study area occurred in the context of the conditions outlined in scenario (1) in which very thick lava flows were rapidly emplaced atop a pre-existing ~1 km thick ice sheet, accumulating to a thickness of ~1.5-1.75 km and causing rapid (several 100 to 10^4 years) and complete top-down melting of the underlying ice sheet.

8. Conclusions

Here we outline the major findings and predictions from our analysis of lava heating and loading of ice sheets in the Late Noachian – Early Hesperian history of Mars.

Initial Lava Flow Emplacement:

Ice sheet lava heating and loading begins with the emplacement of an initial lava flow. We find that the melting rates induced by supraglacial lava flow emplacement and heating are much higher for thinner lava flows, but are sustained for much greater periods of time for thicker lava flows resulting in the melting of much more ice. The meltwater that is produced during the heating process is predicted to infiltrate into the snow and firn layer at the ice sheet surface because the predicted infiltration capacity of this material exceeds even the highest melting rates induced by lava flow emplacement and heating. While the firn layer is predicted to absorb the meltwater produced during initial lava flow emplacement, the surficial snow and firn layer will be very rapidly removed by the heating, melting, refreezing, and compaction associated with lava flow emplacement. Due to these factors the net effect of initial lava flow emplacement will be efficient removal of the surficial snow and firn layer, resulting in the subsidence and degradation of the lava flow. The final degraded initial lava flows will then serve to construct a thermally insulating cap across the ice sheet surface.

If the initial lava flow that is emplaced at the ice sheet surface is very thick (~100 m or greater) then significant, or complete, top-down melting of ice sheets spanning the plausible range of “icy highlands” ice sheet thicknesses (~300-1,000 m) is predicted. As a result of this significant melting, the firn layer has little effect on the meltwater transport and fate since it is removed very quickly in the early stages of lava emplacement and ice sheet heating. Instead, very thick initial lava flows will melt rapidly down into the impermeable ice. In addition, since melting will occur over rapid timescales (several hundred to thousand years), there will not be sufficient time for bottom-up melting of the cryosphere, and thus the subsurface will remain impermeable throughout the lava heating

and loading and top-down melting processes. As a result meltwater is predicted to pool around the flow, where it may then: (1) drain to the ice sheet base forming meltwater reservoirs (since the subsurface will be impermeable) which may source flooding events when the confining ice is ruptured, or (2) drain across the ice sheet surface (if the melt is able to overtop the surrounding ice) in channels following surface topography.

Subsequent Lava Flow Emplacement:

Following the emplacement of the initial lava flow, heat delivery to the underlying ice during the emplacement of subsequent lava flows is delayed after emplacement and significantly reduced in magnitude due to the intervening layers of previously emplaced lava flows. As a result of the reduction in heat delivery, considerably less meltwater is produced throughout the lava flow cooling period. Since the ice sheet surface firn layer is likely to have been removed during initial lava flow emplacement, melting induced by subsequent lava flow emplacement will take place within the impermeable ice. Therefore, meltwater is predicted to pool around the lava flow (possibly inundating the lava flow if enough meltwater is produced) enhancing lava flow cooling and the likelihood of phreatomagmatic events. This meltwater will then either be absorbed by the surrounding unaffected firn layer, or will simply refreeze around the lava flow possibly encapsulating the lava flow in ice. As subsequent lava flows continue to accumulate, the amount of top-down melting induced by heat delivery to the underlying ice will become negligible, and continued accumulation of lava flows will serve chiefly to construct a thermally insulating cap across the ice sheet surface.

If the series of accumulating lava flows are very thick (~100 m or greater), the ice sheet will undergo relatively rapid (several 100 to 10^4 years), complete top-down melting

during the emplacement of only ~1-5 individual lava flows. Therefore very little or no ice sheet may remain to melt during subsequent emplacement. If ice remains, melting will occur at the base of the initial lava flow in the impermeable ice and meltwater will follow the same transport and fate processes as during the initial thick lava flow emplacement (drainage to ice sheet base, or runoff across ice sheet surface).

Continued Lava Heating and Loading:

If the lava flows accumulating at the ice sheet surface are not very thick (~10 m thick or less), the top-down melting induced by direct heat transfer will become negligible as the growing sequence of chilled lava flows prevents downward heat transfer. The sequence of lava flows loaded atop the ice sheet surface will then serve primarily as a thermally insulating layer that will cause the ice-melting isotherm, which would have initially existed at depth below the ice sheet base, to ascend towards the surface. The bottom-up melting processes that result from lava heating and loading in this way can vary depending upon the thicknesses of the individual lava flows being accumulated, the total lava thickness loaded, and the timescale over which lava is accumulated. Due to the highly cratered nature of the topography onto which the Hesperian Planum volcanic plains were emplaced, a dichotomy in lava emplacement conditions is predicted with more rapid and thicker lava accumulations occurring in crater interiors than in the intercrater plains areas. The greater total thicknesses of accumulated lava (~2 km) reached in the crater interiors can allow the ice-melting isotherm to reach the base of the superposed lava flows causing complete thermal instability of the underlying ice sheet, which will then be subject to eventual bottom-up melting. The more rapid lava accumulation (on the order of 10 kyr), will result in bottom-up melting that is limited by

the rate of geothermal heat input from below allowing bottom-up melting of the ice sheet and cryosphere to continue, or even begin, after lava accumulation has completed, leading to “deferred melting.” Conversely, If lava accumulation were to occur gradually (on the order of 100 Myr), then bottom-up melting will be limited by the rate at which insulation is provided by lava accumulation, such that the long-term averaged bottom-up melting rates of the cryosphere and ice sheet will be limited to the rate of lava accumulation. In either case, melting and removal of the ice sheet following lava heating and loading will cause substantial subsidence of the emplaced lava flows. Features predicted to form as a result of this subsidence include: collapse features, depressions, chaos terrain, wrinkle ridges, and fracture systems.

In the intercrater plains, which are predicted to be characterized by thinner accumulations of ice and lava, and by more gradual lava accumulation, insufficient insulation is provided by the thinner total accumulation of lava (~500 m) to allow for the ice-melting isotherm to reach the base of the ice sheet. As a result, no or minimal ice sheet basal melting is predicted in these areas. Given that minimal ice sheet basal melting or subsurface mass loss is predicted in the nominal intercrater plains, few subsidence or collapse related features are predicted to form within the superposed lavas. However, basal melting in the intercrater plains is possible, even at the thinner nominal ice sheet thicknesses (~300 m), if greater thicknesses of lava were accumulated (2 km), or if the mean annual surface temperature and geothermal heat flux were considerably higher (240 K, and 100 mW/m², respectively).

Whether bottom-up melting is limited by the rate of geothermal heat input, or insulation provided by lava accumulation, the bottom-up melting rates of the ice sheets

even at the highest geothermal heat fluxes (100 mW/m^2) are considerably below the infiltration rates predicted for the nominal martian substrate. Therefore, unless there is an extensive layer of very low permeability, or impermeable material, underlying the melting ice sheet, meltwater is predicted to infiltrate into the subsurface to provide groundwater recharge. If a layer of impermeable material (e.g. clay or competent bedrock) underlies the area of bottom-up melting, meltwater may become sequestered at the ice sheet base. Melt sequestered at the ice sheet base would be confined by very large overburden pressures from the superposed ice sheet and lava sequence. This water could be released through ice sheet fracturing near the ice sheet margins resulting in large episodic flooding events.

Cryosphere Implications:

If the cryosphere underlying the ice sheet contains ice, meltwater produced from raising of the ice-melting isotherm during ice sheet lava heating and loading will percolate down into the martian mega-regolith and crust to join any deeper groundwater system. Bottom-up melting of cryospheric ice is nominally predicted to be complete before ice sheet basal melting can initiate. Thus an ice-cemented cryosphere is not predicted to prevent infiltration of meltwater produced at the base of the buried ice sheet.

Effect of Pressure Melting Point Reduction and Ice Impurities:

The analyses we present here have assumed that the ice is free of impurities and that the melting temperature of the ice is constant at 273 K. However, pressure melting point reduction from ice sheet lava heating and loading, and the inclusion of impurities, can allow the ice sheet to undergo basal melting prior to complete melting of the underlying cryosphere by allowing the ice sheet to begin melting at lower temperatures. In this case

the cryosphere would temporarily act as an impermeable barrier, preventing the downward infiltration of melt from the ice sheet, having the same effect as other impermeable formations.

Example Case in Hesperia Planum:

Geological features observed within the study area include fracturing of volcanic crater fill, and a broad (~3 km) fluvial channel, emerging from a volcanically filled crater rim. These features are consistent with predictions made from a lava heating and loading scenario in which the crater was rapidly flooded with very thick lava flows (on the order of 100 m or more), and thus suggest the presence of regional snow and ice deposits in Hesperia Planum during the Late Noachian to Early Hesperian period.

Model Applications:

A wide variety of potential applications exist for the model results and predictions outlined in this study. These include regional mapping to identify locations which may have undergone ice sheet lava heating and loading in order to: (1) Determine to what extent ice sheet lava heating and loading processes may have contributed to the global population of valley networks and outflow channels following the general process outlined in Section 7. (2) Constrain the regional distribution of Late Noachian and Hesperian ice and lava thicknesses through use of the predicted minimum thicknesses required to generate observable ice sheet lava heating and loading morphology. (3) Infer Noachian topography by identifying locations exhibiting the predicted lava-loading subsidence/collapse morphology which are not associated with an obvious impact structure. This may be possible because melting and subsidence are predicted to be minor in intercrater regions, therefore the observation of lava loading morphology in these

locations may indicate the presence of pre-existing underlying topographic depressions which could be reflective of the Noachian surface topography. This would require accurate constraints on Late Noachian and Hesperian snow deposition patterns in order to remove the influence of asymmetrical snow accumulation and ice sheet thicknesses. (4) Constrain Hesperian magma production rates by identifying regional morphological signatures consistent with rapid or gradual lava emplacement.

Acknowledgements

We gratefully acknowledge comments from 2 anonymous reviewers which have helped to improve the quality of the manuscript. We thank David K. Weiss for helpful discussions. We acknowledge support from the NASA Mars Data Analysis Program, Grant NNX11AI81G, and support for participation in the Mars Express High-Resolution Stereo Camera Team (JPL 1237163), both to JWH.

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Lava Flow Thickness (m)	Total Melted (m) (No Firn Layer)	Total Melted (m) (115 m Firn Layer)	Total Melted (m) (17 m Firn Layer)
1	3	8	7
10	30	55	38
100	450	500	458
200	900	956	908

Table 1. The total thicknesses melted from an ice sheet during the emplacement of initial lava flows ranging in thickness from 1-200 m. Total thicknesses melted are shown for the case where no firn layer is present, and for cases where a thick firn layer (115 m) and thin firn layer (17 m) are present. Total melted thicknesses are adjusted for the presence of the firn layers by converting the firn layer into an equivalent thickness of solid ice.

Model Run #	Final Cry. Thickness	Melt Time Start	Melt Time Finished	Final Ice Thickness	Melt Time Start	Melt Time Finished	Final Lava Thickness
1	1,800	51 kyr	1.16 Myr	170	--	--	500
2	440	12 kyr	495 kyr	170	--	--	500
3	0	45 kyr	871 kyr	0	871 kyr	1.08 Myr	2,000
4	0	10 kyr	151 kyr	0	151 kyr	212 kyr	2,000
5	1,800	16.18 Myr	98.34 Myr	170	--	--	500
6	440	10.06 Myr	98.10 Myr	170	--	--	500
7	0	1.67 Myr	81.44 Myr	0	81.44 Myr	85.94 Myr	1,720
8	0	2.57 Myr	39.56 Myr	0	39.56 Myr	44.71 Myr	900
9	1,120	26 kyr	1.19 Myr	870	--	--	500
10	0	12 kyr	112 kyr	610	112 kyr	1.22 Myr	500
11	0	24 kyr	494 kyr	0	494 kyr	1.24 Myr	2,000
12	0	10 kyr	65 kyr	0	10 kyr	331 kyr	2,000
13	1,120	6.25 Myr	98.28 Myr	870	--	--	500
14	0	12.02 Myr	64.12 Myr	610	64.12 Myr	98.30 Myr	500
15	0	1.75 Myr	58.63 Myr	0	58.63 Myr	85.94 Myr	1,720
16	0	3.02 Myr	16.14 Myr	0	16.14 Myr	44.71 Myr	900
17	530	15 kyr	441 kyr	170	--	--	500
18	0	6 kyr	58 kyr	0	58 kyr	268 kyr	500
19	0	13 kyr	152 kyr	0	152 kyr	187 kyr	2,000
20	0	4 kyr	35 kyr	0	35 kyr	56 kyr	2,000
21	530	10.06 Myr	98.08 Myr	170	--	--	500
22	0	12.02 Myr	74.03 Myr	0	74.03 Myr	96.17 Myr	490
23	0	2.56 Myr	41.72 Myr	0	41.72 Myr	46.88 Myr	940
24	0	3.01 Myr	18.53 Myr	0	18.53 Myr	24.19 Myr	490
25	0	10 kyr	156 kyr	700	156 kyr	830 kyr	500
26	0	--	--	0	6 kyr	251 kyr	500
27	0	8 kyr	75 kyr	0	75 kyr	240 kyr	2,000
28	0	--	--	0	5 kyr	91 kyr	2,000
29	0	6.05 Myr	78.14 Myr	700	78.14 Myr	98.24 Myr	500
30	0	--	--	0	0 kyr	92.06 Myr	470
31	0	1.55 Myr	19.64 Myr	0	19.64 Myr	46.88 Myr	940
32	0	--	--	0	0 kyr	23.06 Myr	470

Table 2. Tabulated results from the thermal model runs testing the parameter space illustrated by Figure 8 (-- indicates no bottom-up melting). Results include the corresponding model run number, the final thickness of the cryosphere and buried ice sheet after all melting has taken place, the thickness of lava required to complete melting, and the timescales over which bottom-up melting of the cryosphere and ice sheet occurred (given by times of bottom-up melt initiation and completion).

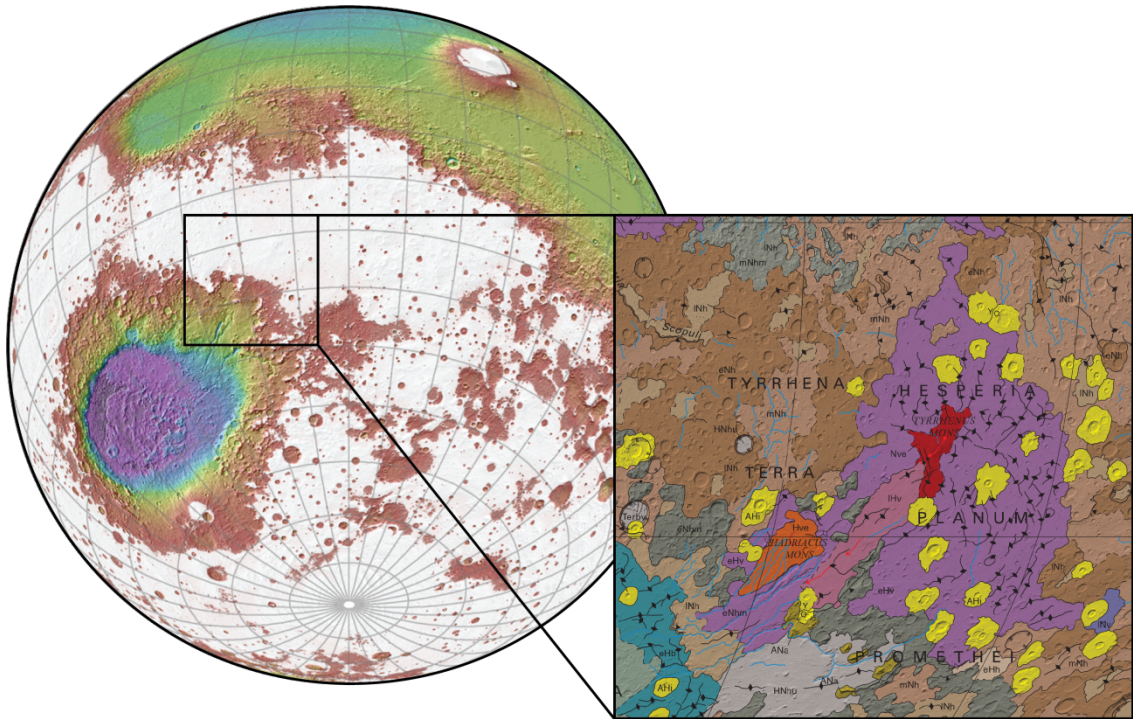


Figure 1. Topographic map depicting a conceptual representation of the “icy highlands” early Mars climate scenario with all regions above the predicted +1 km equilibrium line altitude above which snow and ice are predicted to accumulate shaded white (Head and Marchant, 2014). Inset shows a geological map of the study region, Hesperia Planum (Tanaka et al., 2014b).

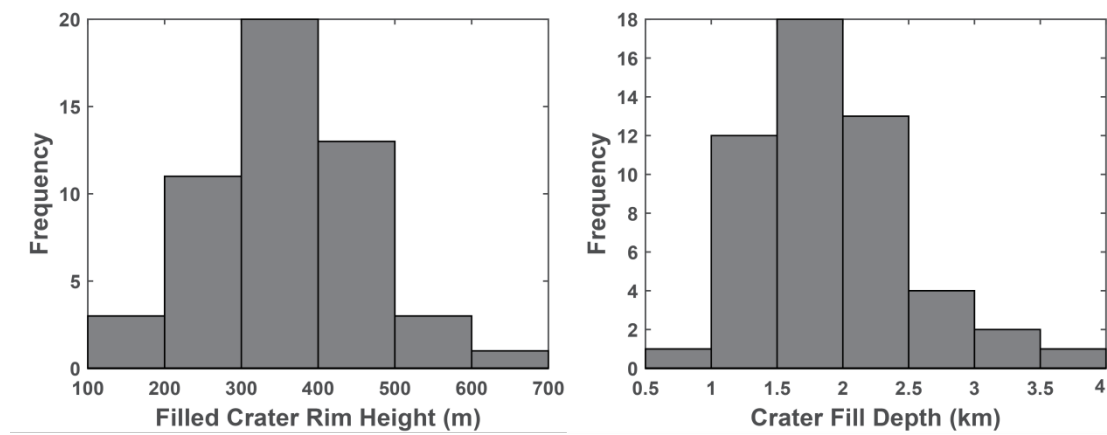


Figure 2. Histograms showing the distribution of crater rim crest height and crater fill depth estimates derived from crater scaling laws (Craddock et al., 1997; Tornabene et al., 2014) using measured crater diameters from buried and partially-buried craters within the Hesperia Planum region.

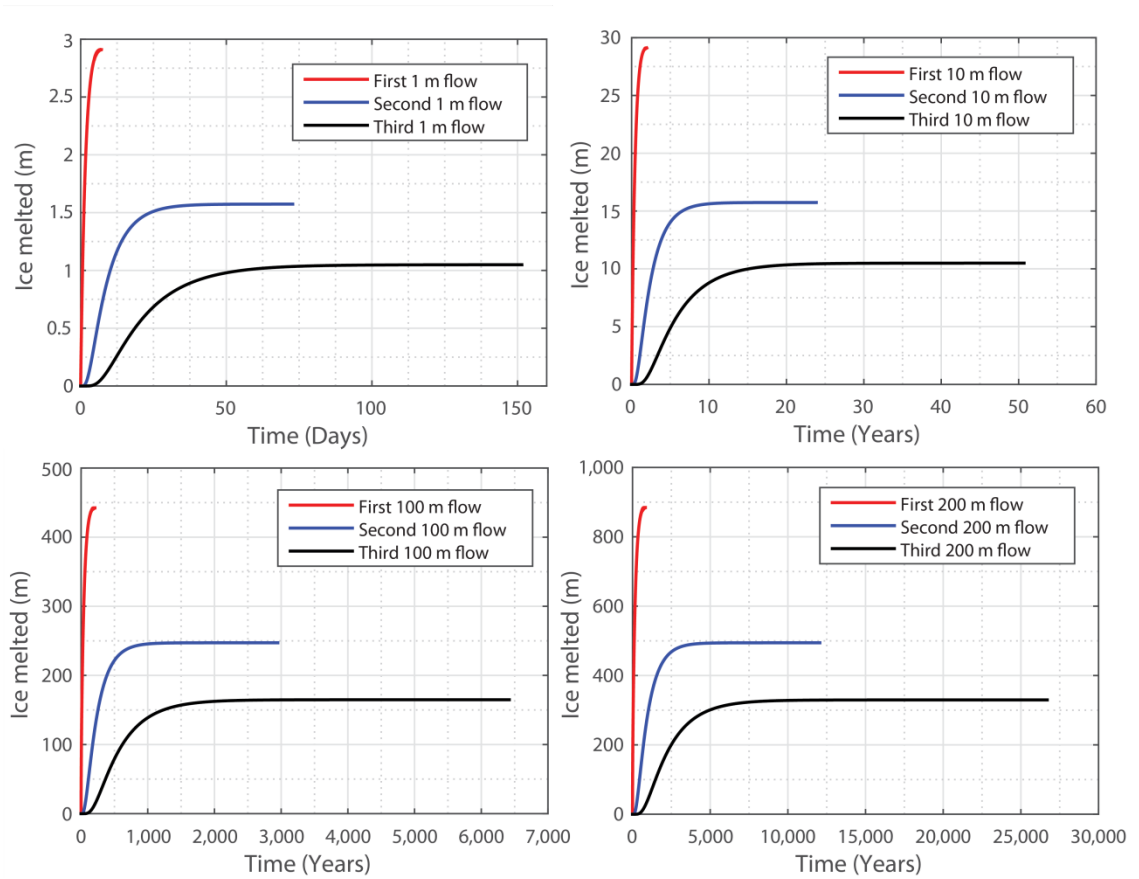
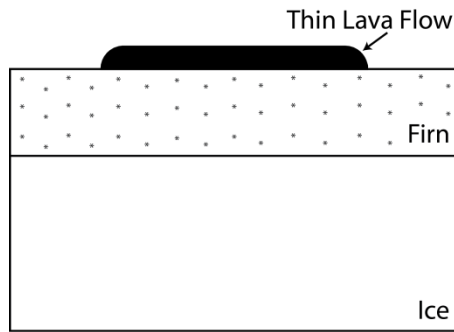
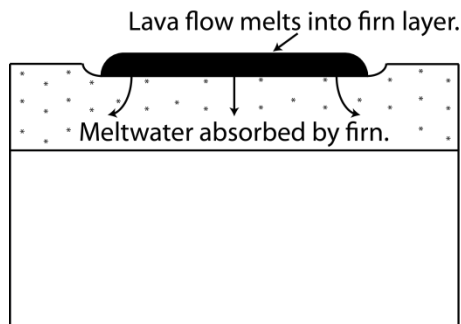


Figure 3. Plots of top-down melting versus time induced by the emplacement of a sequence of lava flows from the initial lava flow (here labeled as first), up to the third lava flow emplaced for all lava flow thicknesses evaluated in this contribution. Heat transfer (and thus ice sheet melting) from subsequent lava flows is delayed and reduced in magnitude due to the intervening presence of previously emplaced lava flows impeding heat delivery to the underlying ice.

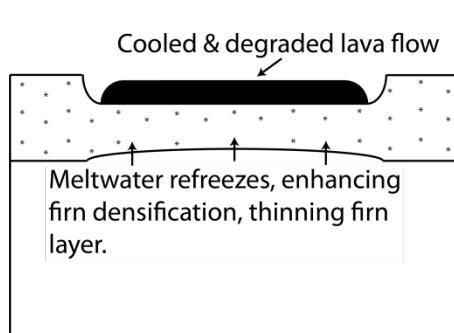
(a.) THIN LAVA FLOW EMPLACED



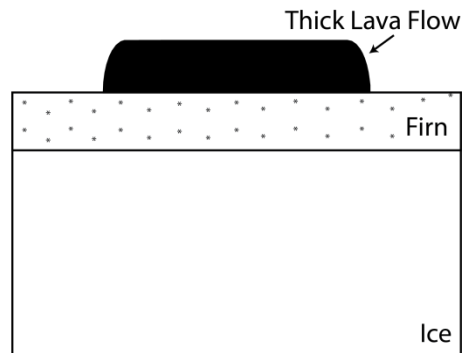
(b.) HEAT TRANSFER & ICE MELTING



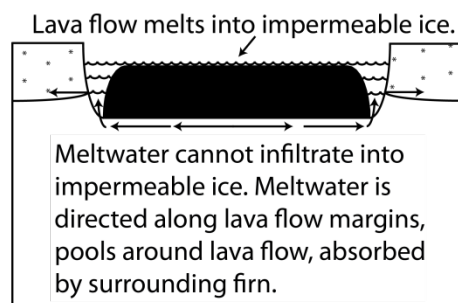
(c.) FOLLOWING LAVA COOLING



(d.) THICK LAVA FLOW EMPLACED



(e.) HEAT TRANSFER & ICE MELTING



(f.) FOLLOWING LAVA COOLING

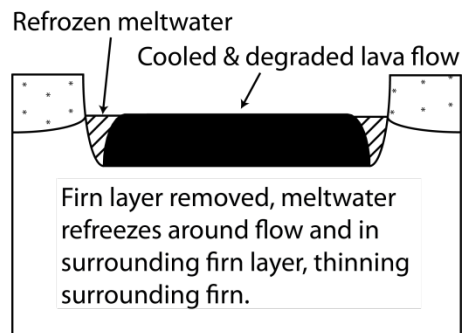


Figure 4. Initial supraglacial lava flow emplacement processes as a function of lava flow and firn layer thickness. In cases where the emplaced lava flow is thin (**a.**), or the firn layer is very thick, the initial lava flow will melt down into the firn layer, and the water that is produced will be absorbed by the firn layer (**b.**). This water will refreeze within the firn layer, enhancing densification, and thus thinning the firn layer (**c.**). If a thick lava flow is emplaced (**d.**), or if the firn layer is thin, then the lava flow will melt down into the impermeable ice. In this case, melt water will be directed along the interface between the lava flow and the ice by the overburden pressure of the lava flow (**d.**). The meltwater will be absorbed by the surrounding firn layer if it is encountered, or will pool around the flow, which will raise the likelihood of phreatomagmatic eruptions and may inundate the flow if enough meltwater is produced (**e.**). If the meltwater is absorbed by the firn layer, it will refreeze in the firn enhancing densification. If the meltwater does not encounter the firn it will refreeze around the lava flow, possibly encapsulating the lava flow in ice if enough melt was pooled (**f.**).

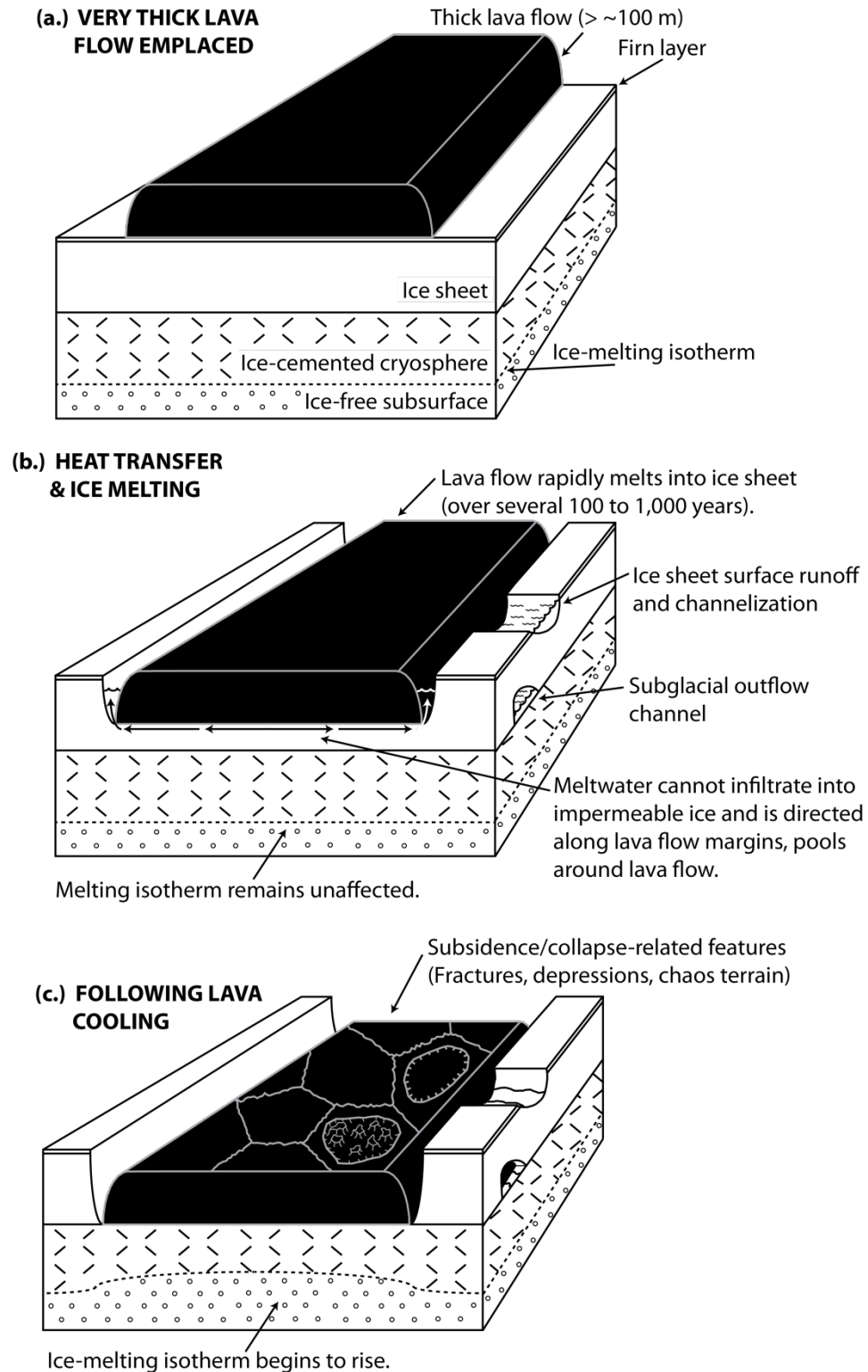
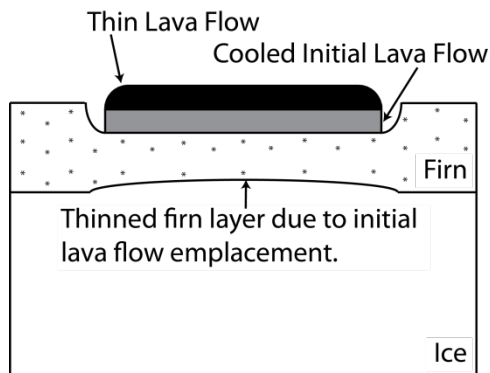


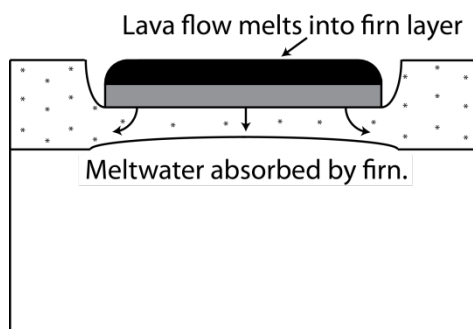
Figure 5. During the emplacement and cooling of very thick lava flows (~100 m or greater) (a.), significant, or complete, top-down melting of ice sheets spanning the plausible range of “icy highlands” ice sheet thicknesses (~300-1,000 m) is predicted. Due to the significant top-down melting, the firn layer will have little effect and the lava flow

will rapidly melt into the impermeable ice sheet **(b.)**. During this process the impermeable ice-cemented cryosphere will remain unaffected because the timescale of top-down melting (several 100 to 10^4 years) is far more rapid than the timescale required for cryosphere reduction and melting (many 10^6 years). Therefore, throughout the melting process, the underlying material will be continuously impermeable, and as a result melt will migrate to the lava flow margins under the overburden pressure of the lava flow **(b.)**. Meltwater will pool around the lava flow, and in topographically favorable directions. The meltwater may then overwhelm or overtop the confining ice, initiating channel formation as it drains toward the glacial margins following ice sheet surface topography **(b.)**. Meltwater may also become trapped at the base of the buried ice sheet, and may fracture the confining ice near the glacial margins, creating subglacial outflow channels and large flooding events **(b.)**. The significant top-down melting resulting from lava flow emplacement will cause an equal amount of subsidence in the superposed lava flows. This subsidence will cause the formation of collapse-related features within the lava flow including fracture systems, depressions, and chaos terrain **(c.)**.

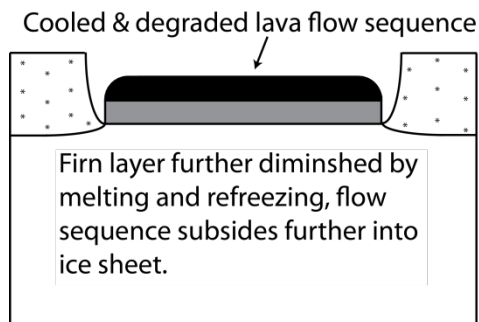
(a.) SUBSEQUENT THIN LAVA FLOW EMPLACED



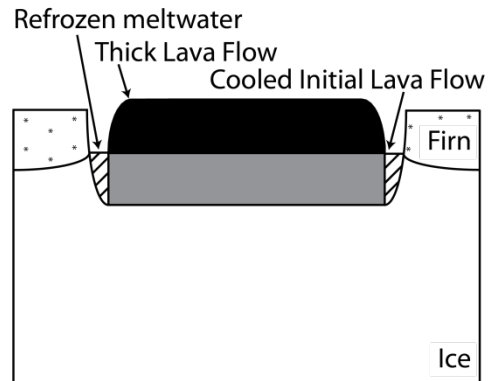
(b.) HEAT TRANSFER & ICE MELTING



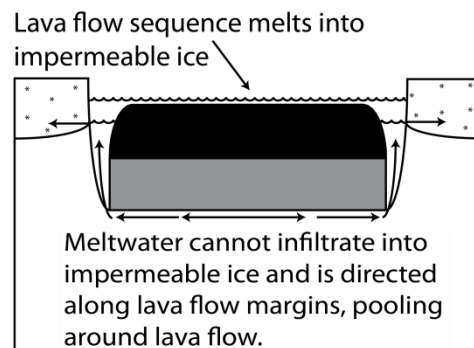
(c.) FOLLOWING LAVA COOLING



(d.) SUBSEQUENT THICK LAVA FLOW EMPLACED



(e.) HEAT TRANSFER & ICE MELTING



(f.) FOLLOWING LAVA COOLING

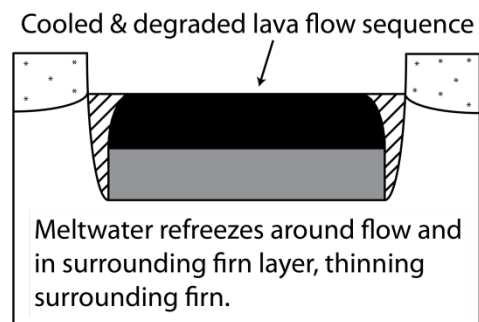
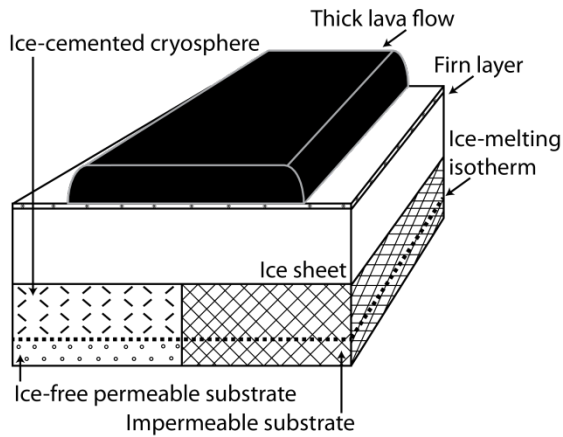


Figure 6. Diagrams illustrating the subsequent supraglacial lava flow emplacement process as a function of lava flow and firn layer thickness. If the emplaced lava flows are thin (a.), or the firn layer is thick, a portion of the firn layer may remain after the emplacement of the initial flow. After a subsequent flow is emplaced atop the initial lava flow, the sequence of lava flows will melt further into the firn layer. Meltwater produced during the cooling of the subsequent flow will be absorbed by the surrounding firn layer (b.). If the lava sequence is able to melt into the impermeable ice, meltwater will be directed along the lava flow margins, where it will pool around the lava flow, or be absorbed by the surrounding firn layer. If the emplaced lava flows are thick (d.), or the firn layer is thin, then the initial lava flow will have already melted into the impermeable ice. As a result, meltwater produced by subsequent lava flow emplacement will be directed along the lava flow margins by the overburden pressure of the lava sequence (e.).

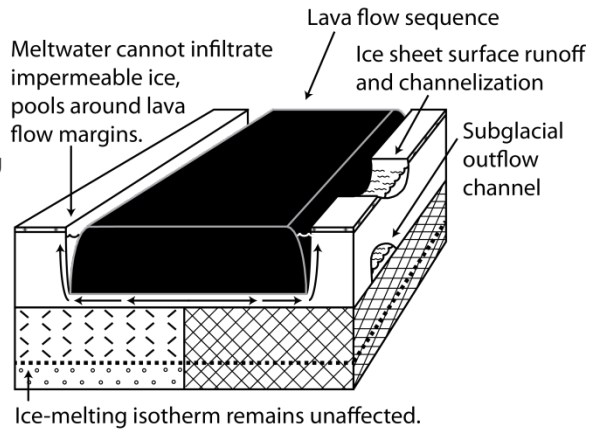
The meltwater will pool around the flow sequence (**e.**), enhancing cooling and raising the likelihood of phreatomagmatic events and possibly inundating the flow sequence if enough melt is produced. If the meltwater is able to rise far enough it may be absorbed by the surrounding firn layer (**e.**), otherwise it will refreeze around the lava flow possibly encapsulating the lava flow in ice if enough melt was pooled (**f.**).

RELATIVELY THICK LAVA FLOW EMPLACEMENT

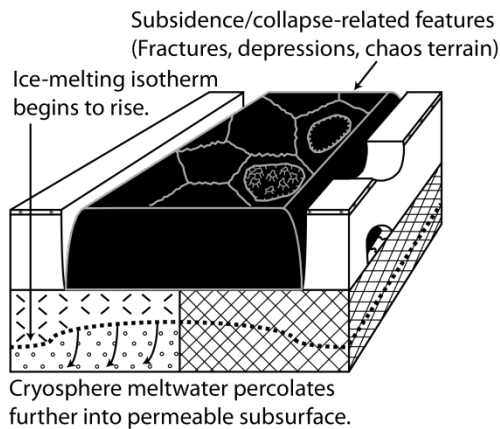
(a.) INITIAL LAVA EMLPACEMENT



(b.) HEAT TRANSFER & TOP-DOWN MELTING



(c.) BOTTOM-UP MELTING



(d.) FOLLOWING COOLING & MELTING

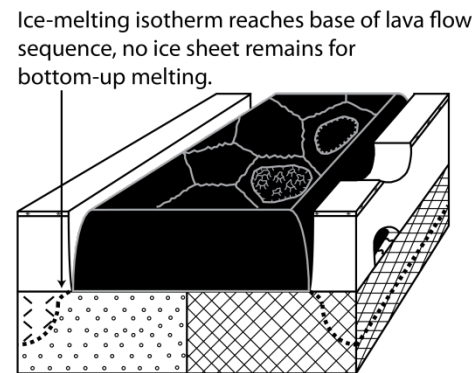


Figure 7. Synthesized illustration of the ice sheet lava heating and loading processes in the case where very thick lava flows (~ 100 m thick or greater) are accumulated at the ice sheet surface. In this scenario, the supraglacial emplacement of very thick (~ 100 m or greater) lava flows causes rapid (several 100 to 10^4 yr) and complete melting of the ice sheets by top-down melting (a., b., and c.) due to the large amount of heat energy stored in the lava flows and the efficient transfer of that energy to the ice. Due to the significant top-down melting produced in this scenario, the firn layer will be very rapidly removed, having a negligible effect on meltwater transport. The lava flow will quickly melt down into the impermeable ice, causing meltwater produced at the lava-ice interface to migrate along the margins of the lava flow, pooling around the flow and in topographically favorable directions (b.). The meltwater may then overwhelm or overtop the confining ice, initiating channel formation as it drains toward the glacial margins following ice sheet topography (b.). During this process the impermeable ice-cemented cryosphere will remain unaffected because the timescale of top-down melting (several 100 to 10^4 yr) is far more rapid than the timescale required for cryosphere reduction and melting (many 10^6 years). As a result, meltwater may also become trapped at the base of the buried ice

sheet, and may fracture the confining ice near the glacial margins, creating subglacial outflow channels and large flooding events **(b.)**. Top-down melting of the subjacent ice sheet resulting from lava flow emplacement will cause subsidence of the superposed lava flows equal in amount to the thickness of the melted ice sheet. This potentially significant subsidence will cause the formation of collapse-related features within the lava flow including fracture systems, depressions, and chaos terrain **(c.)**. Following top-down melting, the insulation provided by the large thickness of emplaced lava flows will cause the ice-melting isotherm to rise towards the surface. This will result in bottom-up cryosphere melting, liberating meltwater from the subsurface pore-ice that will percolate further into the subsurface providing groundwater recharge **(c.)**. The ice-melting isotherm may eventually reach the base of the lava flow sequence, however, no ice sheet basal melting will take place because the ice sheet will have been removed by top-down melting **(d.)**

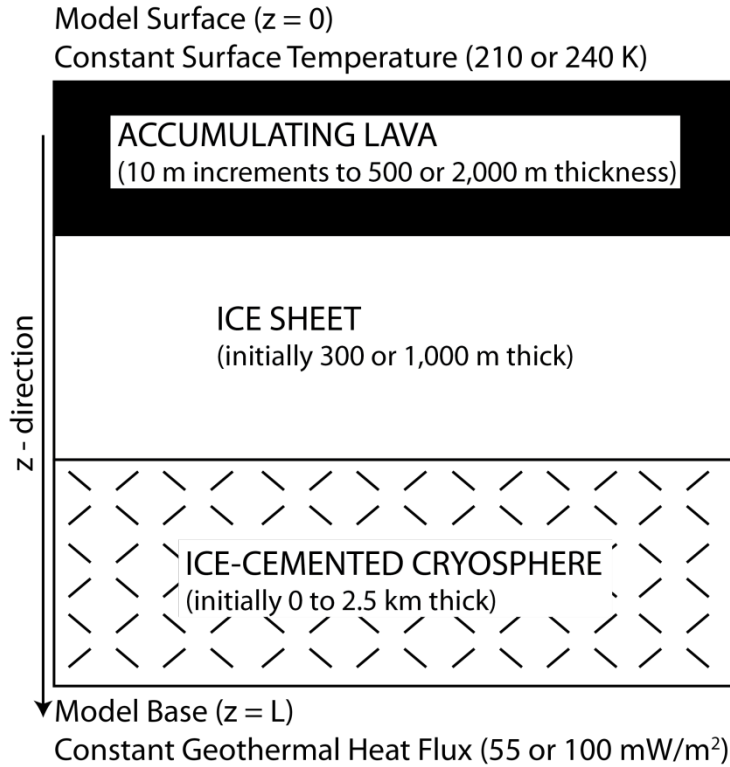


Figure 8. Conceptual illustration of the numerical thermal model described in section 5.1-5.7. The thermal model domain spans the depth from the surface of the accumulating lavas ($z = 0$), through the buried ice sheet, and down to the initial depth of the cryosphere ($z = L$). At the top of the model, a constant temperature is held at either 210 or 240 K based on the end-member estimates for the Late Noachian mean annual surface temperature established in section 5.1. At the base of the model a constant geothermal heat flux is held at either 55 or 100 mW/m², based on end-member scenarios for the regional geothermal heat flux established in section 5.1. Initial ice sheet thicknesses are set to either 300 or 1,000 m (testing the end-member cases of the Late Noachian surface water inventory as discussed in section 5.2), and the initial cryosphere thickness is set based on the depth of the ice-melting isotherm under the combination of the surface temperature, ice sheet thickness, and geothermal heat flux. Lavas are then accumulated in 10 m increments up to a maximum depth of 500 or 2,000 m and the induced melting in the cryosphere and ice sheet is tracked through time.

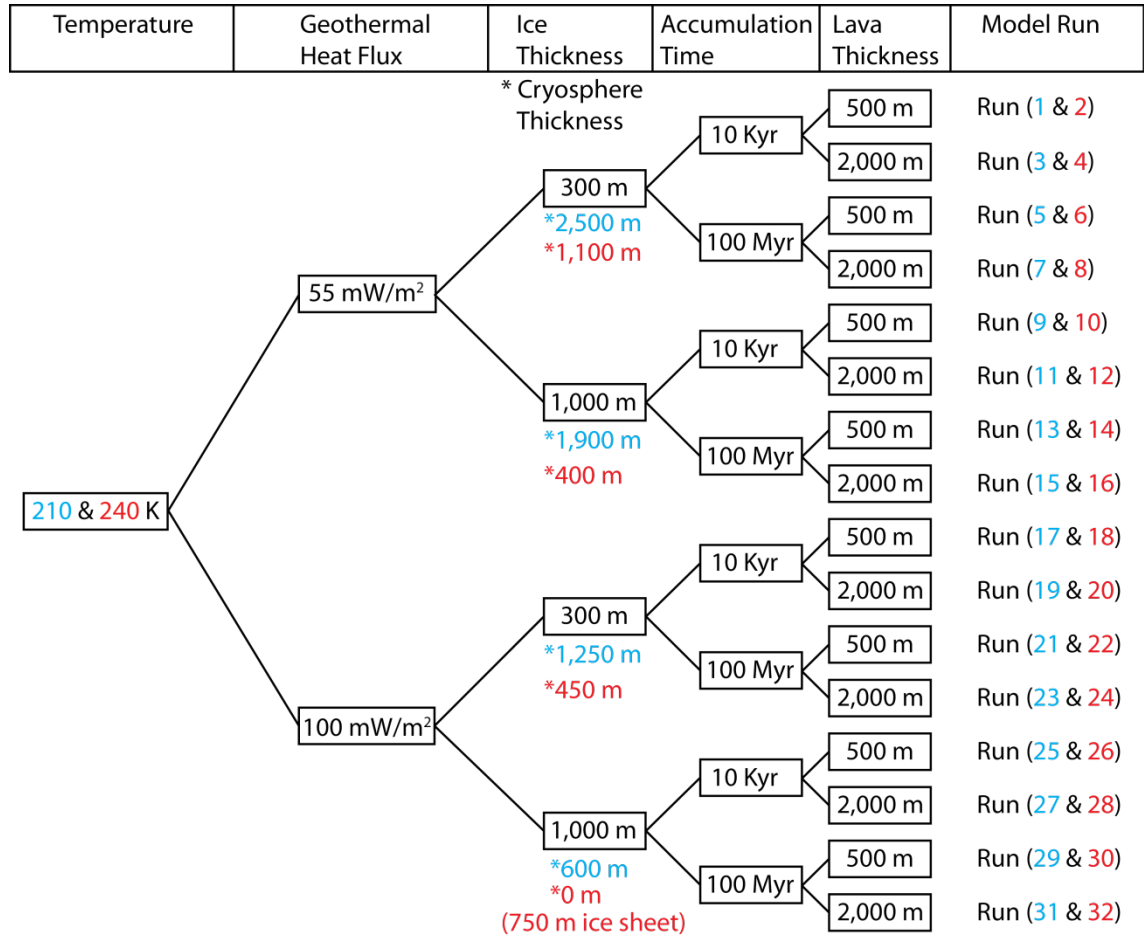


Figure 9. Schematic diagram illustrating the parameter space explored with the thermal model in the assessment of the ice sheet lava heating and loading process. The numbers marked with an asterisk underneath the ice sheet thickness correspond to the initial cryosphere thickness as determined by the combination of the surface temperature, geothermal heat flux, and ice sheet thickness. The red-colored cryosphere thicknesses and model run numbers correspond to the 240 K surface temperature cases, while the blue-colored cryosphere thicknesses and model run numbers correspond to the 210 K surface temperature cases.

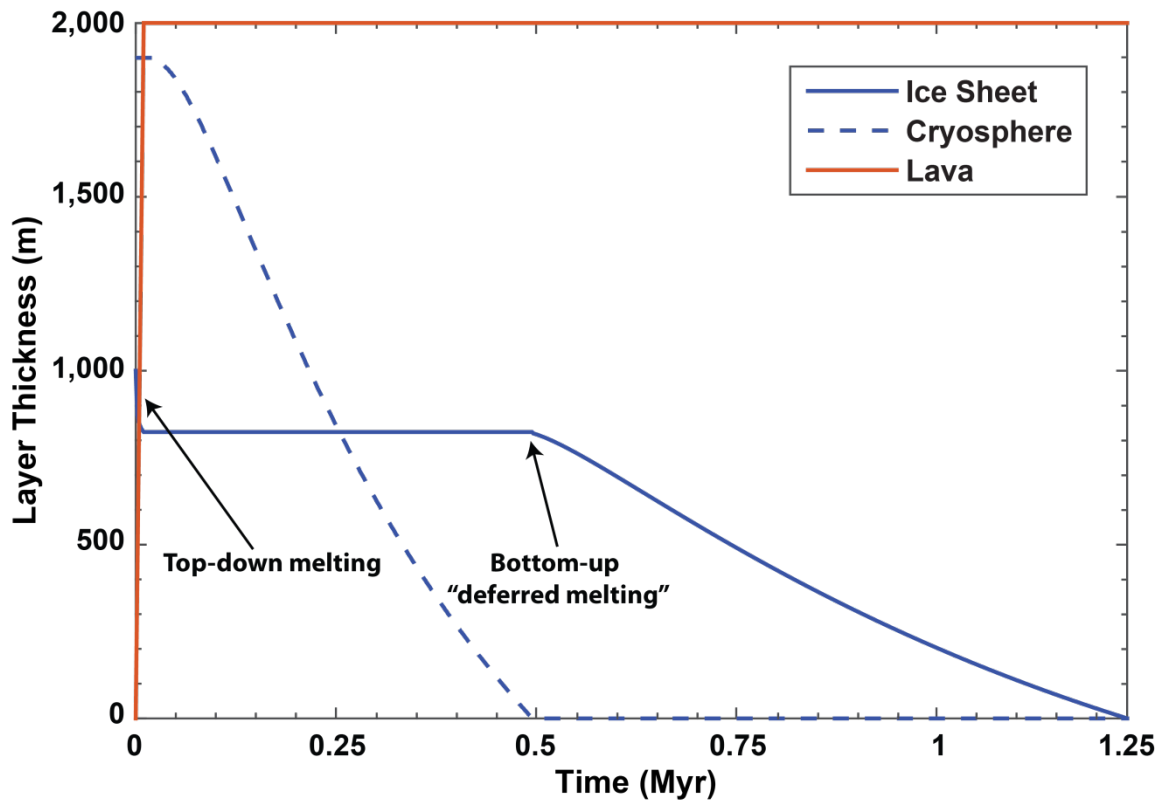


Figure 10. Thermal model output showing results from the nominal crater interior lava heating and loading scenario (model run #11; Table 2). In this scenario, a 2 km sequence of lava is rapidly loaded in 10 kyr atop the ice sheet surface. During emplacement, top-down melting causes an initial sharp decrease in the ice sheet thickness. This process defers the initiation of bottom-up melting of the cryosphere by ~24 kyr (Table 2) by reducing the total amount of insulation. After top-down melting becomes negligible, the insulating effect of the superposed lava becomes dominant, and the ice-melting isotherm begins to ascend toward the surface. In this case, lava loading completes over shorter timescales than bottom-up melting can respond. Thus, bottom-up melting of the cryosphere does not take place until ~14 kyr after the lava has completed accumulating, while bottom-up melting of the ice sheet does not begin until ~500 kyr after lava accumulation has completed, leading to “deferred melting.” Due to the large amount of insulation provided by the 2 km thick lava sequence, complete bottom-up melting of the cryosphere then proceeds over a timescale of ~0.5 Myr followed by complete bottom-up melting of the ice sheet over ~0.75 Myr. The bottom-up melting rates of the cryosphere and ice sheet in this scenario are governed by the rate at which heat is input into the system by the geothermal flux.

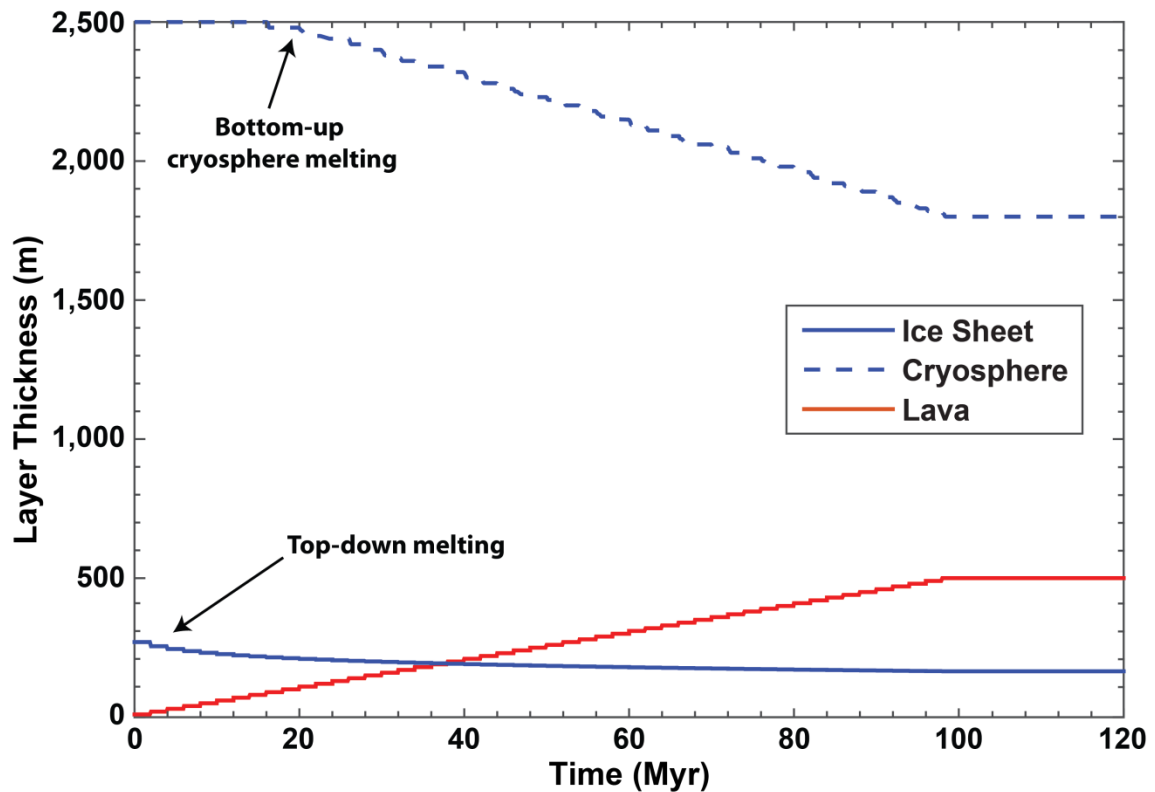
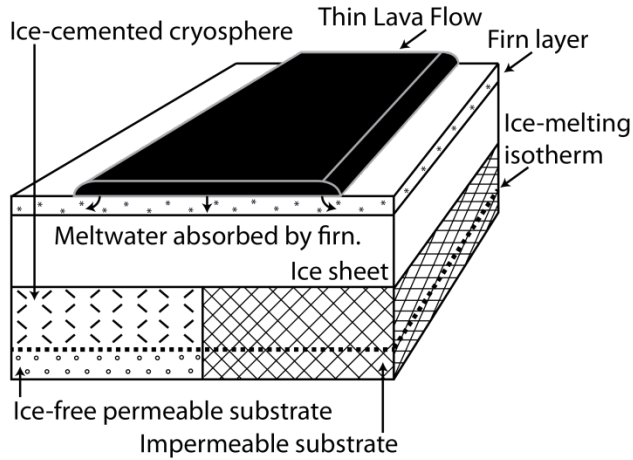


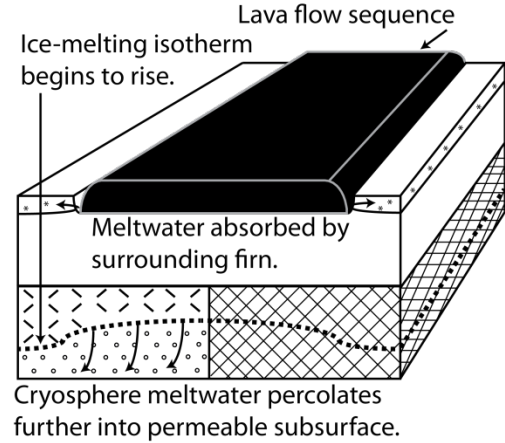
Figure 11. Thermal model output showing results from the nominal intercrater plains lava heating and loading scenario (model run #5; Table 2). In this scenario, a 500 m sequence of lava is loaded atop the ice sheet surface over the course of 100 Myr. The larger amounts of top-down melting which occur earlier in the lava emplacement process cause a relatively sharp initial decrease in ice sheet thickness. The reduction in total insulation across the model resulting from this process defers the onset of bottom-up melting of the cryosphere by ~16 Myr (Table 2). After top-down melting becomes negligible, the insulating effect of the superposed lava becomes dominant, and the ice-melting isotherm begins to ascend toward the surface. In this scenario, lava loading occurs more slowly than the associated bottom-up melting processes. As a result, after each increment of lava is accumulated, the ice-melting isotherm rises by an incremental amount and bottom-up melting proceeds (at a rate governed by the geothermal heat input) to the new equilibrium depth of the ice-melting isotherm before the next increment of lava is accumulated. Due to this process, the long-term averaged bottom-up melting rates are limited to the rate of lava accumulation. In this scenario, the insulation provided by the 500 m thick lava sequence is only sufficient to raise the ice-melting isotherm by ~700 m, such that the cryosphere remains after the lava heating and loading process at a thickness of ~1.8 km and no ice sheet basal melting takes place.

RELATIVELY THIN LAVA FLOW EMPLACEMENT

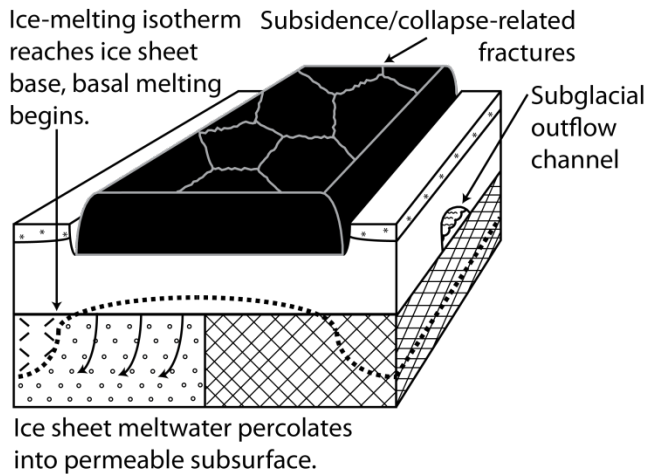
(a.) INITIAL LAVA EMLPACEMENT



(b.) CONTINUED LAVA ACCUMULATION



(c.) BOTTOM-UP ICE SHEET MELTING



(d.) FOLLOWING COOLING & MELTING

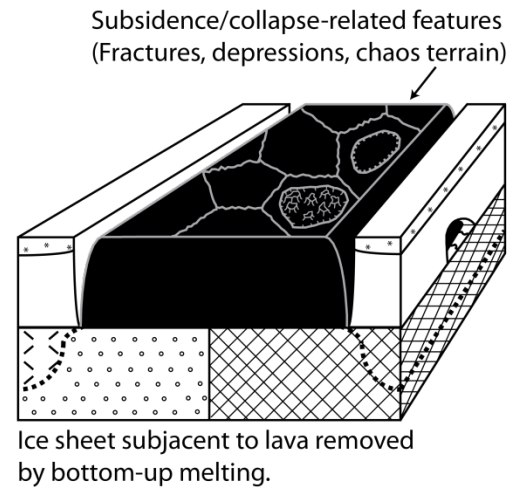


Figure 12. Synthesized illustration of the ice sheet lava heating and loading processes in the case where relatively thin lava flows (~10 m thick or less) are accumulated at the ice sheet surface. In this case top-down melting is quickly limited as accumulation proceeds due to the growing sequence of chilled lava flows which prevent downward heat transfer from freshly emplaced lava flows to the underlying ice sheet. Meltwater produced by top-down melting in this scenario is predicted to be predominantly absorbed by the firn layer which lies across the surface of the ice sheet (a.) subduing ice sheet surface runoff. The meltwater will refreeze within the firn, causing further densification and eventually removing the effected firn layer in combination with direct melting and compaction (b.). As the thickness of the superposed lava flow sequence increases, the amount of top-down melting induced by continued lava emplacement will become negligible. At this point the insulating effect of the superposed lava flow sequence will become dominant and will begin to lift the ice-melting isotherm towards the surface (b.), initiating bottom-up cryospheric melting (meltwater produced from this will percolate further into the

subsurface). In the nominal intercrater plains scenario, the lava flows are not predicted to accumulate to thicknesses sufficient to raise the ice-melting isotherm up to the base of the ice sheet, and thus the final result of the intercrater plains lava heating and loading scenario is represented by panel **(b.)**. If lavas were to accumulate to greater thicknesses, the insulation provided would continue to lift the isotherm, which would eventually intercept the ice sheet base resulting in ice sheet basal melting **(c.)**. Bottom-up melting rates are ultimately limited by the heat input from the geothermal heat flux, and are predicted to be substantially less than the infiltration capacity of the martian subsurface. As a result meltwater is predicted to infiltrate into the subsurface to provide groundwater recharge **(c.)**. However, if the substrate is intrinsically impermeable (e.g. due to a clay layer), then meltwater will pool at the ice sheet base, and may fracture the confining ice near the ice sheet margins leading to large flooding events **(c.)**. If the superposed lava flows reach a sufficient thickness to raise the ice-melting isotherm to the ice sheet base (~2 km), then the entire buried ice sheet will be eventually removed by bottom-up melting **(d.)**. Melting and removal of the buried ice sheet will cause subsidence of the superposed lava sequence which will result in the formation of collapse features, depressions, chaos terrain, and fractures which will be expressed at the surface. If the same thickness of lavas were accumulated more rapidly (~10 kyr), then the ice-melting isotherm would not be able to rise to the lava sequence base before accumulation completed. In this case, “deferred bottom-up melting” of the ice sheet would occur, otherwise resulting in the same processes and geological features as the gradual accumulation case.

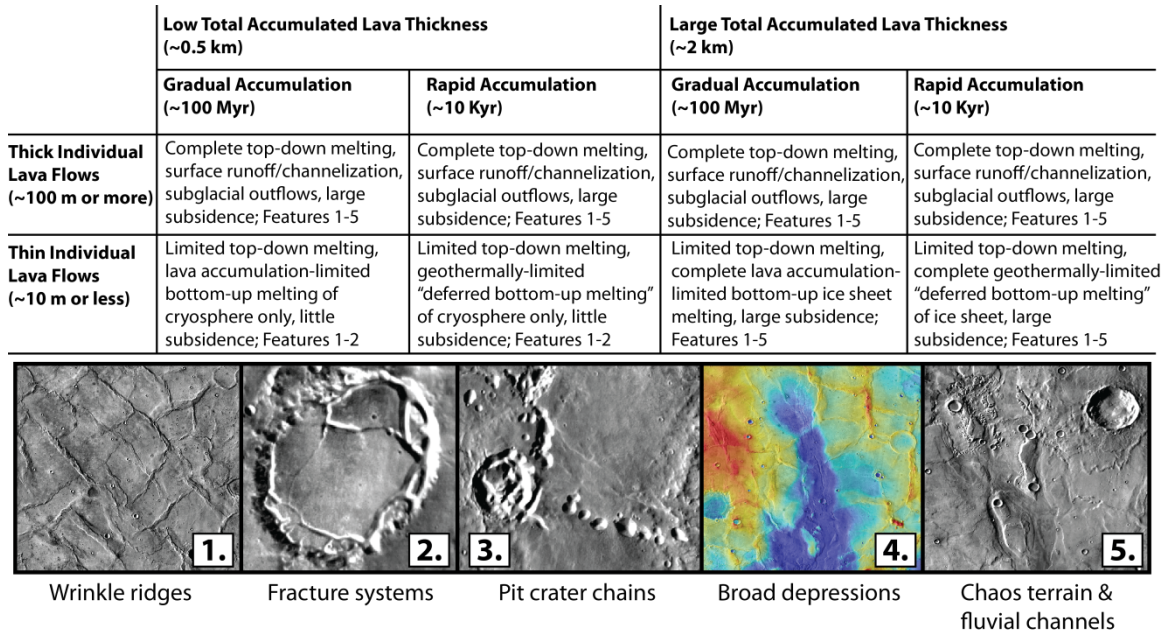


Figure 13. Characteristic processes predicted to occur, and geological features predicted to form, as a result of the ice sheet lava heating and loading process as a function of individual lava flow thickness (thick lava flows ~100 m or greater, and thin lava flows ~10 m or less), total accumulated lava flow thickness (low total accumulated lava thickness of ~0.5 km, and a large total accumulated lava thickness of ~2 km), and lava accumulation timescale (gradual lava accumulation over ~100 Myr, and rapid lava accumulation over ~10 kyr). Potential examples of each of the predicted geological features within the Hesperia Planum region are documented in Thermal Emission Imaging System (THEMIS) global daytime infrared imagery (Christensen et al., 2004) and shaded Mars Orbiter Laser Altimeter (MOLA) data (Smith et al., 2001). Features include: (1) wrinkle ridges, (2) fracture systems, (3) pit crater chains, (4) broad depressions, and (5) chaos terrain, and fluvial channels. While the geological features contained in panel 5 are noted in the predictions of the lava heating and loading scenarios with thin individual lava flows (~10 m or less) and a high total accumulated lava thickness (~2 km), fluvial channels are only predicted to form if the substrate underlying the buried ice sheet is impermeable. In this case meltwater can pool below the ice sheet and rupture confining materials near the glacial margins causing flooding events and channel formation. Otherwise, if the underlying material is permeable, the meltwater produced in these two scenarios is generated to slowly to exceed the infiltration capacity of the Mars crustal materials and is predicted to infiltrate into the subsurface to provide groundwater recharge.

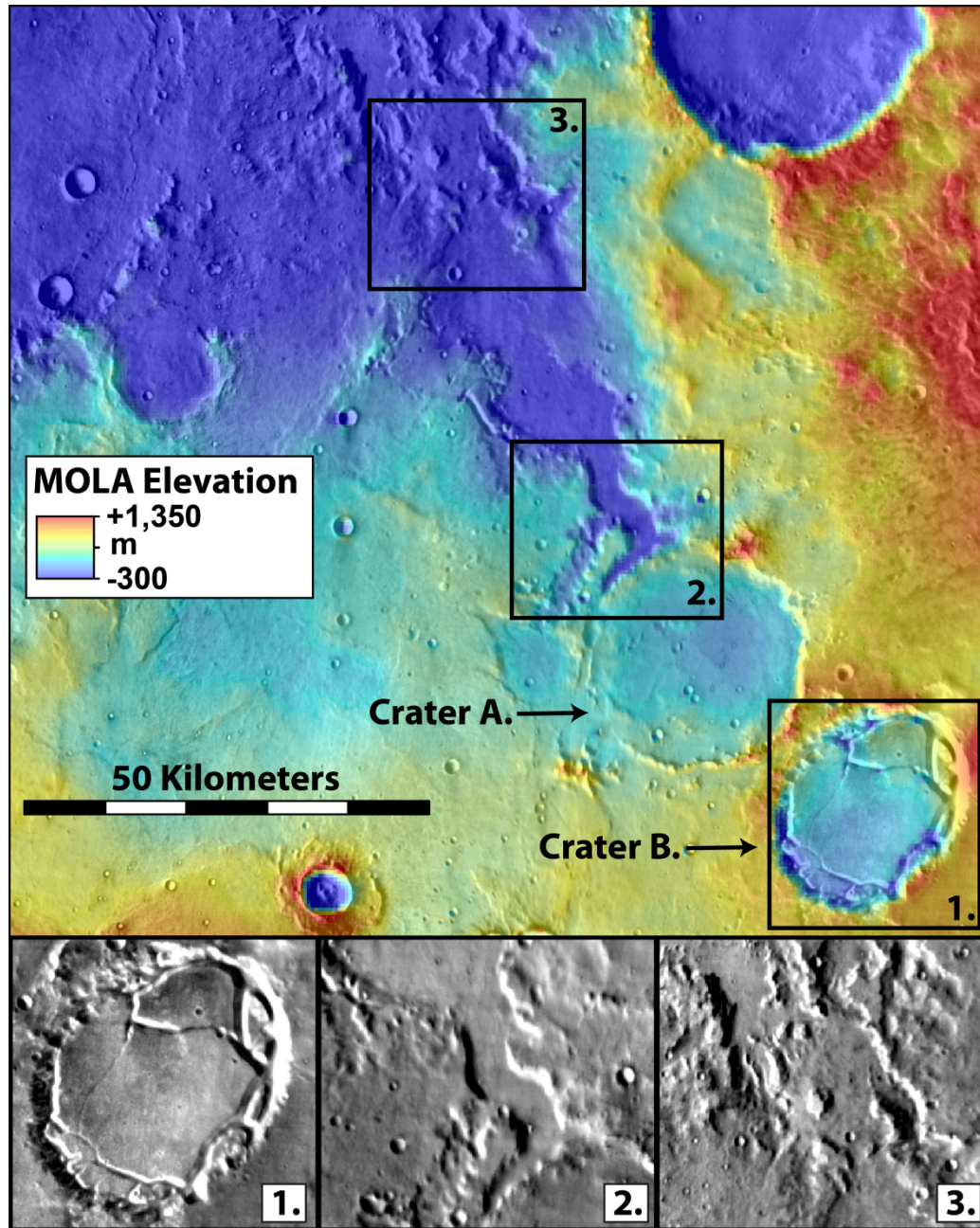


Figure 14. Context image of the selected study area in northern Hesperia Planum (centered approximately at 106.496 °E and 6.205 °S) examined in section 7. Base imagery is composed of THEMIS global daytime infrared data (Christensen et al., 2004), with overlain shaded MOLA elevation data (Smith et al., 2001) measured relative to the Mars datum. Craters A and B, discussed in the text, are labeled. Inset images highlight features in the study area predicted to occur as a result of lava heating and loading. These are (1) highly fractured volcanic crater fill, (2) a channel emerging from the rim of crater A, and (3) tear-drop shaped islands appearing in the floor of the channel suggesting a fluvial origin.

**Chapter 2: Formation of Outflow Channels on Mars: Testing the Origin
of Reull Vallis in Hesperia Planum by Large-scale Lava-Ice Interactions
and Top-down Melting**

James P. Cassanelli^a & James W. Head^a

^a Department of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

Published in: *Icarus*

Cassanelli, J. P., and Head, J.W., (2018). Formation of outflow channels on Mars:

Testing the origin of Reull Vallis in Hesperia Planum by large-scale lava-ice interactions

and top-down melting. *Icarus*, 305, 56-79, doi: 10.1016/j.icarus.2018.01.001

Abstract

The Reull Vallis outflow channel is a segmented system of fluvial valleys which originates from the volcanic plains of the Hesperia Planum region of Mars. Explanation of the formation of the Reull Vallis outflow channel by canonical catastrophic groundwater release models faces difficulties with generating sufficient hydraulic head, requiring unreasonably high aquifer permeability, and from limited recharge sources. Recent work has proposed that large-scale lava-ice interactions could serve as an alternative mechanism for outflow channel formation on the basis of predictions of regional ice sheet formation in areas that also underwent extensive contemporaneous volcanic resurfacing. Here we assess in detail the potential formation of outflow channels by large-scale lava-ice interactions through an applied case study of the Reull Vallis outflow channel system, selected for its close association with the effusive volcanic plains of the Hesperia Planum region. We first review the geomorphology of the Reull Vallis system to outline criteria that must be met by the proposed formation mechanism. We then assess local and regional lava heating and loading conditions and generate model predictions for the formation of Reull Vallis to test against the outlined geomorphic criteria. We find that successive events of large-scale lava-ice interactions that melt ice deposits, which then undergo re-deposition due to climatic mechanisms, best explains the observed geomorphic criteria, offering improvements over previously proposed formation models, particularly in the ability to supply adequate volumes of water.

1. Introduction

Much work has been dedicated to understanding the mechanisms responsible for the formation of the outflow channels on Mars, enormous channels carved into the surface of Mars during the Hesperian and Amazonian periods (Baker and Milton, 1974; Baker, 1982a; Baker et al., 1992; Carr, 1979; M. H. Carr, 1996; Carr, 2002; Clifford, 1993; Head et al., 2003; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007). Formation of the outflow channels has canonically been interpreted to have occurred by catastrophic eruptions of groundwater from pressurized confined aquifers (Baker and Milton, 1974; Baker, 1982a; Baker et al., 1992; Carr, 1979; Carr, 1996; Carr, 2002; Clifford, 1993; Head et al., 2003; Harrison and Grimm, 2004; Andrews-Hanna and Phillips, 2007; Harrison and Grimm, 2008) resulting from excess hydraulic head generated by elevated recharge (Clifford, 1993; Harrison and Grimm, 2004; Russell and Head, 2007) or from the downward propagation of a confining cryosphere layer from cooling of the climate (Carr, 1979; Carr, 1996; Carr, 2002; Hanna and Phillips, 2005; Wang et al., 2006).

These models for outflow channel formation, however, have been challenged by difficulties in generating sufficient hydraulic head (Carr, 2002) and overpressure (Hanna and Phillips, 2005), requiring very high aquifer permeability (Head et al., 2003; Manga, 2004; Harrison and Grimm, 2008), and by limited recharge sources (Russell and Head, 2007; Harrison and Grimm, 2009; Cassanelli et al., 2015). Given the difficulties with the canonical outflow channel formation models, many alternative candidate formation mechanisms have been proposed, including: (1) Melting of cryospheric ground-ice by extrusive and intrusive volcanic activity (Squyres et al., 1987). (2) Subglacial volcanism and near-surface intrusive activity (Chapman and Tanaka, 2002). (3) Melting of buried ice deposits (Zegers et al., 2010; Roda et al., 2014). (4) Breaching and release of surficial

water reservoirs (Clifford and Parker, 2001; Coleman and Baker, 2007; Harrison and Grimm, 2008; Coleman and Baker, 2009). (5) Dewatering of sedimentary materials (Montgomery and Gillespie, 2005). (6) The collapse of subsurface water conduits (Marra et al., 2014; Rodriguez et al., 2015; Rodriguez et al., 2016).

In light of climate model results predicting the accumulation of regional ice sheets across the martian highlands in early Mars history (Forget et al., 2013; Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth et al., 2015; Wordsworth, 2016), and the widespread distribution of contemporaneous volcanism on Mars (Head et al., 2002; Crown et al., 2007; Gregg and Crown, 2009; Werner, 2009; Carr and Head, 2010; Robbins et al., 2011; Tanaka et al., 2014) (Figure 1), recent work has proposed that large-scale lava-ice interactions could serve as a potential alternative mechanism for outflow channel formation (Cassanelli and Head, 2016). Here we assess in detail the potential formation of outflow channels by large-scale lava-ice interactions through an applied case study of the Reull Vallis outflow channel system, selected for its close association with the effusive volcanic plains of the Hesperia Planum region (Crown et al., 1992; Crown and Greeley, 1993; Ivanov et al., 2005; Crown et al., 2007; Gregg and Crown, 2009; Williams et al., 2009; Mest et al., 2012; Tanaka et al., 2014b) (Figure 2 and 3). We first review the geomorphology of the Reull Vallis system (Figure 3) to outline criteria that must be met by the proposed formation mechanism. We then assess local and regional lava heating and loading conditions and generate model predictions for the formation of Reull Vallis to test against the outlined geomorphic criteria.

2. The Reull Vallis Outflow Channel System

The Reull Vallis outflow channel system is a segmented network of fluvial valleys

(Mest and Crown, 2001; Mest and Crown, 2002; Mest and Crown, 2003; Ivanov et al., 2005; Kostama et al., 2007; Mest and Crown, 2014) in the eastern rim region of the Hellas basin (Figure 2). The system consists of 5 individual segments (Kostama et al., 2007) which stretch over a distance of ~1,700 km westward from Hesperia Planum over Promethei Terra and down toward the Hellas basin interior (Figure 3). These individual segments are grouped into the upper and lower Reull Vallis on the basis of distinct channel morphologies, formation times, and water sources (Kostama et al., 2007). The upper portion of the Reull Vallis fluvial system (Figure 3) consists of the northern trough, Waikato Vallis (formerly referred to as Segment 1; Mest and Crown, 2001; Ivanov et al., 2005; Kostama et al., 2007; Kostama et al., 2013), the Morpheos basin (bounded by the 650 m contour with the approximate outline shown in Figure 3; Kostama et al., 2007), and Segment 2, while Segment 3 and Teviot Vallis comprise the lower Reull Vallis system (Kostama et al., 2007).

The geomorphology of the Reull Vallis system suggests a relatively complex fluvial history, with multiple episodes of flooding activity (Kostama et al., 2007; Kostama et al., 2013). The upper Reull Vallis system is thought to have formed by episodic drainage from the northern trough through Waikato Vallis (Figure 3 and 4) into the Morpheos basin where water was collected and then released to form the remaining downstream segment (Segment 2) of the upper Reull Vallis system (Figure 3) (Kostama et al., 2013).

The channels of the lower Reull Vallis fluvial system (Segment 3 and Teviot Vallis; Figure 3) are substantially wider and deeper than those contained in the upper Reull Vallis system, and do not contain the hydrodynamically-shaped/streamlined islands, terraces, or secondary branches which characterize the upper system (Kostama et al.,

2007). These morphometric differences have been suggested to indicate that the upper portion of Reull Vallis served as a tributary to the lower system and therefore, that formation of the lower portion of the system occurred by other mechanisms, with different sources of water, and predated that of the upper system (Kostama et al., 2007).

Here, we focus on the upper Reull Vallis system (Figure 3 and 4) due to the close spatial and temporal association between the upper system and the Hesperian Ridged plains unit. In order to outline criteria constraining the formation of the upper Reull Vallis system we now examine the geomorphology of the system in more detail.

3. Upper Reull Vallis Geomorphology

The upper Reull Vallis outflow channel system emanates from the northern trough, a ~140 km long, tens of kilometers wide, ~100-200 m deep linear depression in the south-central region of Hesperia Planum (Figure 3 and 4) (Kostama et al., 2007). The northern trough is predominantly a topographic feature (Kostama et al., 2007) which interrupts the surrounding, relatively flat-lying, Hesperian plains wrinkle ridge unit (Kreslavsky and Head, 1999; Kreslavsky and Head, 2000) and contains a surface which is morphologically indistinguishable from the surrounding wrinkle ridge plains (excluding the presence of several narrow, ~1-3 km wide, sinuous channels concentrated along the base of the trough) (Figure 4). South of the northern trough, the upper Reull Vallis system emerges abruptly as a large steep-sided canyon (Kostama et al., 2007) within Waikato Vallis (formerly Segment 1 of Reull Vallis; Mest and Crown, 2001; Kostama et al., 2007) (Figure 3 and 4).

The Waikato Vallis segment of the upper Reull Vallis system shows an abundance of evidence consistent with an origin by fluvial activity and erosion (Kostama et al., 2007).

This evidence includes the presence of scour marks and streamlined flat-topped islands within Waikato Vallis (Figure 4) which exhibit terracing that is unpaired with terraces observed on the canyon walls, supporting a fluvial erosional origin (Kostama et al., 2007). The terraces within the valley have been interpreted as evidence for multiple phases of flooding activity (Kostama et al., 2007), but could also have potentially formed as a result of step-wise decreases in flow discharge. In addition to the features indicative of fluvial erosion, numerous flat-topped blocks and knobs of Hesperian wrinkle ridge material (Figure 4) occur along the walls of Waikato Vallis which are dissected by fractures extending from the canyon outward into the surrounding wrinkle ridge plateau. The presence of these features is indicative of collapse of the surrounding Hesperia Planum wrinkle ridge plateau inward toward the valley, suggestive of undercutting by subsurface flow (Kostama et al., 2007). Toward the valley interior the blocks and knobs of wrinkle ridge material become increasingly eroded and rounded (Figure 4), showing streamlined and terraced features indicative of greater exposure to flow contained within the valley (Kostama et al., 2007). While it is also possible for collapse and subsidence to have been generated through other processes (such as the movement of magma), the generation of collapse and subsidence within Waikato Vallis by subsurface water flow is more consistent with the geologic context of the system and is supported by the observed suite of geologic features. Therefore we adopt the interpretation of collapse and subsidence caused by undercutting from subsurface water flow as suggested by the geomorphology and advocated by previous investigations.

The abrupt emergence, lack of tributaries, depressed channel head, and collapsed features at the channel head have been interpreted to suggest that the water which carved

the upper Reull Vallis system originated from the subsurface of Hesperia Planum, within the Northern trough (Figure 3 and 4) (Kostama et al., 2007; Capitoli and Mest, 2010). In this interpretation, the Hesperian ridged volcanic plains predate the Reull Vallis outflow channel as well as the northern trough, with the wrinkle ridge morphology of the surface retained through the later formation of the northern trough depression and the outflow channel system. This interpretation is supported in part by the clear erosion of wrinkle ridges which extend into the northern trough basin by the emergence of the outflow channel system in Waikato Vallis. However, the genetic link between the channel and the northern trough is interpretational in nature and it is possible that the northern trough may predate the volcanic plains. In this case the trough may have been filled with volcanic plains during ridged plains emplacement (with contemporaneous formation of wrinkle ridges within and outside the basin) with later deformation of both the trough and plains by channel formation. Here we adopt the interpretation of Kostama et al. (2007) and Capitoli and Mest (2010) with the northern trough serving as the source of the Reull Vallis outflow channel as a consistent plausible scenario to test, noting the interpretation-dependent nature of this sequence of events. Based on this adopted interpretation, the volume of material eroded in the formation of the upper Reull Vallis system is then equal to the total volume of the segments downstream from the Northern trough source region (where the missing volume is due to the expulsion of subsurface water as opposed to eroded surface materials). Therefore, the eroded volume of the upper Reull Vallis system is equal to the sum of the volumes of Waikato Vallis and Segment 2, approximately $2,200 \text{ km}^3$ (Kostama et al., 2007). Assuming an average substrate porosity of 0.3 (Clifford, 1993; Kleinhans, 2005) this total eroded volume yields an eroded sediment

volume of $\sim 1,500 \text{ km}^3$.

In order to estimate the volume of water required to erode this observed valley sediment volume, we adopt an upper-limit volumetric sediment load of 0.4 (the maximum load which can be carried before the flood transitions to a debris flow, a value significantly higher than those typical of terrestrial rivers which are on the order of 0.001; Andrews-Hanna and Phillips, 2007), which yields a minimum water volume of $\sim 2,300 \text{ km}^3$ required to produce the observed erosion in the upper Reull Vallis system.

These minimum volume estimates are based exclusively on the eroded volume of the downstream segments and assume that no extra water was required to fill the Morpheos basin during valley formation and that the volume of the valley segments has not been altered since formation (either increased by erosion or decreased by infilling). The latter assumption is likely to be invalid because the system has clearly undergone post-formation erosion and because Segment 2 of the upper Reull Vallis system contains evidence of lineated valley fill (Mest and Crown, 2001b; Baker, 2001; Mest and Crown, 2002; Mest and Crown, 2003; Mest and Crown, 2014). Lineated valley fill has been shown to be formed by debris-covered glaciers (Head et al., 2006a; Head et al., 2006b; Head et al., 2010), a process that can lead to valley widening and shallowing by erosion and infilling which would result in post-formation modification of the valley volumes. However, our minimum volume estimates are retained in order to assess a conservative minimum water volume required for the formation of the upper Reull Vallis and because the volume of the lineated valley fill (Levy et al., 2014) is undetermined in detail in this region.

Finally, we estimate the fluid flux required to sustain flow within the upper Reull

Vallis system. In order to make this estimate, we implement the Darcy-Weisbach equation in an approach advocated by Wilson et al. (2004) for the assessment of flow conditions in martian channels, with discharge given by:

$$Q = d W \left(\frac{8 g h_R S}{f_c} \right)^{1/2} \quad (1)$$

where Q is the discharge (m^3/s), d is mean flow depth (m), W is the mean channel width (m), g is gravitational acceleration (3.71 m/s^2 on Mars), h_R is the hydraulic radius (m), S is the sine of the down-flow slope of the channel bed, and f_c is the dimensionless Darcy-Weisbach friction factor which depends on the nature of the channel bed and the ratio of the flow depth to the channel bed roughness scale.

While the Darcy-Weisbach equation may yield the most accurate martian channel flux estimates, the approach still suffers from a number of shortcomings in application to outflow channels. These include: (1) dependence upon the poorly-constrained flow depth, channel bed roughness, and Reynolds number, and (2) prediction of zero or negative discharge for level or upslope sections of channel (these factors are discussed in greater detail by Wilson et al., 2004 and Andrews-Hanna and Phillips, 2007). Despite the potential inaccuracies involved, the approximate flux estimates yielded by this approach are sufficient for the purposes of this analysis.

Flux estimates are generated through the use of channel width measurements taken from high-resolution CTX images, channel depths derived from individual Mars Orbiter Laser Altimeter (MOLA) elevation points, and channel slopes from fits of gridded MOLA elevation profiles along the channel floors (as illustrated in Figure 5). In order to produce approximately average flux estimates we assume a gravel bed (intermediate

between sand and boulder beds) channel to compute the dimensionless friction factor involved in the Darcy-Weisbach equation (Wilson et al., 2004). The influence of variations in the channel bed material, and the other remaining parameters, on the discharge estimates produced through the Darcy-Weisbach method are later explored in detail. The flux estimates produced through this analysis represent maximum end-member values because bankfull flow conditions are assumed in the absence of interior channel dimensions or evidence for paleo-water levels within Reull Vallis. The bankfull flow assumption necessarily overestimates the flow discharge because the channel bed is inherently assumed to have remained fixed at the post-flow elevation throughout the flooding event (again assuming that no material has been added to the valley floor since formation). We later examine the water depths that would be produced within the system at lower volume fluxes.

Measurements have been performed at several locations (Figure 3-5) along both Waikato Vallis and Segment 2 to provide an inclusive range of flux estimates. The measured valley dimensions and flux estimates derived from this analysis of the upper Reull Vallis system are summarized in Table 1, with location numbers corresponding to the points indicated in Figures 3-5. Location 3 represents a branching point of Waikato Vallis, thus at this location measurements were collected from each branch of the valley (locations 3a and 3b) and summed to produce a total flux estimate. We adopt the flux estimate derived from locations 3a and 3b as the most conservative flux estimate for the upper Reull Vallis system, as flow terraces are preserved at this location (Kostama et al., 2007) and flux estimates have been produced from measurements (width, depth, slope) of the inner-most terraces at these locations (Figure 5).

An important unknown is the nature of the substrate into which the Reull Vallis outflow channel was carved which can have implications for the water discharge rates, water volumes, and timescales required for channel formation. We earlier assumed that the upper Reull Vallis outflow channel was formed within a colluvium substrate in order to estimate the water fluxes and volumes required to form the channel. However, the precise nature of the substrate into which the upper Reull Vallis channel incised is unclear and therefore requires further investigation.

Estimates of the volume of water, and therefore the timescale, required for channel formation can be affected by the assumed substrate nature, with canonical wisdom suggesting that incision and erosion into a bedrock substrate requires significantly greater water volumes and erosion timescales relative to that in a regolith substrate. Under bedrock substrate conditions, therefore, the volumetric sediment concentration within the flow would be decreased, with a corresponding increase in the ratio of required water volume to eroded valley volume. This, however, has been called into question by recent observations of rapid bedrock canyon formation through efficient erosion by plucking and abrasion (Lamb and Fongstad, 2010; Sweeney and Roering, 2017), with erosion rates potentially limited by the ability of the flow to transport sediment (Lamb and Fongstad, 2010). Nevertheless, in the volume analysis performed here, a highly conservative volumetric sediment load of 0.4 was adopted to yield a minimum required volume of water, likely a substantial overestimate of the sediment load regardless of substrate nature. Therefore, we maintain our minimum required water volume estimate and reemphasize that the volume of water involved in the formation of the Reull Vallis outflow channel was likely to have been greater than our derived estimate, particularly if

formation occurred within a bedrock substrate.

The effect of the substrate nature on the discharge estimate method utilized in the volume analysis performed here is manifest primarily through the variation of channel bed roughness and flow resistance. Wilson et al., (2004) explored the effect of a range of substrate natures (sand bed, gravel bed, boulder bed) on channel discharge estimates through variations in the Darcy-Weisbach friction factor and found that, under martian conditions, differing channel bed conditions produced no more than a 25% variation in discharge estimates for any given channel dimension. This result was attributed to the large water depths typical of Mars channels in relation to the bed roughness scales (Wilson et al., 2004). We have derived discharge estimates by adopting a representative average gravel bed friction factor, as the gravel bed is characterized by a bed roughness scale intermediate between fine alluvium and boulders. Any effect of the substrate nature on the discharge estimates produced are well within the margin of error introduced by uncertainties in the paleo flow depth and are therefore neglected (though we explore the sensitivity of our discharge estimates to these parameters below). While the nature of the substrate in which the Reull Vallis outflow channel formed is unclear, our discharge and volume estimates are largely unaffected. Further constraints on the substrate nature could allow for more accurate discharge estimates and for more realistic, less conservative, estimates of the flow sediment load and required water volume.

In order to evaluate the potential range in discharge estimates for the upper Reull Vallis system a sensitivity analysis has been performed accounting for a conservative measurement error and an expanded parameter space. Discharge estimates have been calculated using equation (1) for the upper Reull Vallis at location 3 (Figures 3-5)

assuming a width and depth measurement error fraction of ± 0.2 (Figure 6a), and for a range in bed friction factors (Figure 6b and 6c), and slopes (Figure 6d and 6e). The minimum bed friction factor value is computed assuming a sand bed with a D_{50} of 0.001 m (corresponding to a medium sand) while the maximum is computed assuming a boulder bed with a D_{84} of 0.25 m (approximately equal to the average D_{90} value measured from the Viking 1, Viking 2, and Pathfinder landing sites; Wilson et al., 2004). The lowest slope measured from the Reull Vallis system in this analysis, 0.15° , is adopted as the minimum slope value, while the maximum slope is set to 1° , approximately double the largest slope directly measured (Table 1). The sensitivity analysis (Figure 6) illustrates the dependence of discharge upon channel width, depth, friction factor and slope with discharge increasing with increasing width, depth, slope, and decreasing friction factor. The sensitivity analysis (Figure 6) indicates that for the range of measurement error considered, and the plausible range of input parameters, discharge estimates do not vary from the nominal estimate ($10^7 \text{ m}^3/\text{s}$) by more than an order of magnitude, with minimum and maximum estimates of $\sim 10^6 \text{ m}^3/\text{s}$ and $2 \times 10^7 \text{ m}^3/\text{s}$, respectively.

The geomorphologic characteristics of the upper Reull Vallis system outlined above place several important constraints on the nature of any potential formation mechanisms. These constraints are listed below, with more detailed explanations of their interpretational origins and relative uncertainties discussed in the preceding text.

(1) Formation post-dating Hesperian ridged plains emplacement (as evidenced by the deformation and erosion of the wrinkle ridge plains near the channel head, and the stratigraphic superposition of the channel downstream).

(2) Episodic release of water, consistent with observations suggesting multiple periods of flooding activity (suggested by the presence of terraces within Waikato Vallis and Segment 2; Figures 3-5).

(3) Subsurface water flow or undercutting (as indicated by the subsidence and collapse features in the northern trough and Waikato Vallis; Figures 3 and 4).

(4) Subsidence of ~100-200 m within the northern trough channel head, with retention of the Hesperian ridged plains surface morphology (evidenced by the morphology of the northern trough).

(5) Production and release of a minimum of ~2,300 km³ of water (equal to the minimum amount of water needed to erode the volumes of the upper Reull Vallis segments).

(6) Supply of a volumetric flux of water of ~1x10⁷ m³/s (indicated by flux estimates derived from the most conservative measurements available for the upper Reull Vallis system).

With the morphologic constraints on the origin of the upper Reull Vallis system outlined, we now explore the potential genetic relationship between Hesperia Planum volcanism and the upper Reull Vallis fluvial system by assessing whether or not large-scale lava-ice interactions and ice sheet lava heating and loading (*e.g.* Cassanelli and Head, 2016) can account for the formation of the valley system.

4. Formation of the Upper Reull Vallis System: Relation to Hesperia Planum Volcanism

The interaction of ice sheets and ice-cemented regolith with magmatic processes (extrusive volcanism, intrusive dike and sill formation) has been explored quantitatively on the Earth and Mars (*e.g.* Squyres et al., 1987; Wilson and Head, 2002; Head and

Wilson, 2002; Wilson and Head, 2007; Head and Wilson, 2007; Edwards et al., 2012; Dundas and Keszthelyi, 2013; Wilson et al., 2013). The relationship between Hesperia Planum volcanism (Greeley and Crown, 1990; Crown and Greeley, 1993; Crown et al., 1992; Gregg et al., 1998; Gregg et al., 2002; Gregg and Crown, 2005; Goudy et al., 2005; Crown et al., 2007; Gregg and Crown, 2007; Williams et al., 2007; Williams et al., 2008; Gregg and de Silva, 2009; Williams et al., 2009; Williams et al., 2010; Rampey and Harvey, 2012; Gregg, 2017) and outflow channel development has previously been explored by several investigators (Baker, 1982a; Squyres et al., 1987; Crown et al., 1992; McKenzie and Nimmo, 1999; Tanaka et al., 2002; Chapman and Tanaka, 2002) who proposed heating and melting of ground-ice through extrusive and intrusive volcanism as a possible outflow channel formation mechanism (Figure 7a). In these formation models, melting of ground-ice caused by extrusive and intrusive volcanic heating is suggested to result in the subsurface flow of fluidized regolith material, removing the structural support of the superposed Hesperian ridged plains and causing subsidence and collapse (Baker, 1982; Squyres et al., 1987; Crown et al., 1992) (Figure 7a). Outflow channels are then incised by the down slope discharge of meltwater to the surface through steady or episodic discharge after collecting in the subsurface (Squyres et al., 1987). This model has been applied specifically to the Dao and Harmakhis outflow channel systems (Squyres et al., 1987) which originate in the southwest region of Hesperia Planum. These channel systems exhibit morphologies that are broadly similar to those of the upper Reull Vallis system, with fluvial channels extending from collapsed source regions in Hesperian wrinkle ridge units.

While the formation model invoking volcanic melting of ground-ice (Squyres et al.,

1987) is generally able to account for the distinctive morphology of the upper Reull Vallis fluvial system, with the collapsed source region and downstream fluvial incision (Figure 3 and 4), several difficulties remain. (1) The subsided volume of the Northern trough source region, $\sim 426 \text{ km}^3$ (Kostama et al., 2007), is substantially less than that of the incised fluvial valleys ($\sim 2,200 \text{ km}^3$; Kostama et al., 2007) and the minimum volume of water required to carve the valleys ($\sim 2,300 \text{ km}^3$). Therefore, melting and expulsion of ground-ice equivalent to the subsided volume of the source region alone cannot provide a sufficient quantity of water to produce the erosion observed in the upper Reull Vallis system. (2) Heating and melting of ground-ice by intrusive magmatism, favored over extrusive heating (Squyres et al., 1987), would cause bottom-up melting and therefore preferentially result in the migration of meltwater deeper into the subsurface (Cassanelli et al., 2015; Cassanelli and Head, 2016). Thus additional mechanisms or conditions would be required to cause discharge of meltwater to the surface to produce channel incision. (3) Ground-ice melting and regolith fluidization would produce flow with inherently low water to sediment ratios, approximately equal to the porosity of the fluidized regolith which estimates suggest is generally less than ~ 0.3 (Clifford, 1993; Hanna and Phillips, 2005; Kleinhans, 2005). Under these conditions, the flow would be characterized as a debris flow, with sediment fractions of ~ 0.7 or higher. Kleinhans (2005) has cast doubt on the ability of debris flows to have formed the large martian outflow channels because expected run-out lengths are much less than those of the channels. Thus, additional mechanisms or conditions would be required to account for separation of the meltwater from the regolith sediment load to produce a fluvial flow with higher water to sediment ratios.

Given the inconsistencies between the predictions of the volcanic ground-ice melting model and the morphology of the upper Reull Vallis, we now explore the ice sheet lava heating and loading model (Cassanelli and Head, 2016) as an alternative formation mechanism.

4.1 Formation of the Upper Reull Vallis System: Ice Sheet Lava Heating and Loading

The ice sheet lava heating and loading model (Figure 7b) (Cassanelli and Head, 2016) proposes the emplacement of effusive lava flows atop regional ice sheets predicted to exist in the martian highlands in early Mars history (Forget et al., 2013; Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth et al., 2015) by widespread contemporaneous effusive volcanic activity (*e.g.* Head et al., 2002).

Emplacement of lava flows atop the regional ice sheets could have occurred by (1) direct advancement of lava flows onto the ice from topographically-elevated terrain (as has been observed in glaciated terrestrial volcanic settings; Edwards et al., 2012; Edwards et al., 2013; Edwards et al., 2014; Edwards et al., 2015), (2) supraglacial eruptions fed by dikes emplaced through the ice sheet due to high strain rates (Wilson and Head, 2002; Head and Wilson, 2002; Head and Wilson, 2007; Wilson and Head, 2007) or, (3) eruptions from tuyas that have pierced through the ice sheet (*e.g.* Smellie, 2007). Lava-ice interactions could have occurred in Hesperia Planum in geometries which did not involve supraglacial lava flow emplacement. However, this particular lava-ice interaction geometry provides a useful test of the volcano-ice origin for the Reull Vallis outflow channel because this geometry optimizes heat transfer from the lava flows to the ice (with the exception of subglacial/englacial sill formation, though melting by this mechanism would be limited by the areal extent of the sill and could produce less total melting).

Therefore, by considering supraglacial lava flow emplacement, we estimate maximum melting rates and volumes which can be produced by lava-ice interactions.

Ice sheet lava heating and loading has been suggested as a potential outflow channel formation mechanism due to predictions (Cassanelli and Head, 2016) of significant meltwater production and subsequent collapse and deformation of the superposed lava flows from meltwater evacuation and loss of structural support (Figure 7b). Melting by ice sheet lava heating and loading can be produced in a top-down mode by direct contact heat transfer from the lava (Figure 8), and in a bottom-up mode through the formation of an insulating layer causing more efficient confinement of geothermal heat within the ice (Figure 9) (Cassanelli and Head, 2016).

The timescales of top-down melting are much more rapid ($\sim 10^0$ - 10^3 yr) than those of bottom-up melting ($\sim 10^6$ - 10^8 yr), with the melting rates increased accordingly (Cassanelli and Head, 2016). Both the top-down and bottom-up melting processes are influenced by the thermophysical state of the ice sheet. Under typical Late Noachian-Early Hesperian conditions (e.g. Forget et al., 2013; Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth et al., 2015; Wordsworth, 2016), ice sheet structure is predicted (Cassanelli and Head, 2015) to be characterized by a relatively thin layer of snow and firn (with thickness dependent upon the total thickness of the ice sheet, ice accumulation rate, ice sheet age, and surface temperatures) underlain by ice of increasing density and a layer of ice-cemented substrate (the cryosphere) that lies above the ice-melting isotherm. The snow and firn layer acts to reduce the density and thermal conductivity, increasing the susceptibility of the ice sheet to melting under both top-down and bottom-up heating conditions (Cassanelli and Head, 2015).

The amount and rate of top-down ice sheet melting generated by lava heating and loading depends upon the thickness of the emplaced lava flows and upon the thickness of any preexisting lava deposits, with thicker lava flows in more direct contact with the ice producing more melting at greater sustained melting rates (Figure 8c and 8d). If lava loading occurs by the accumulation of thin lava flows, top-down ice sheet heating and melting is predicted to be less efficient (Cassanelli and Head, 2016), leading instead to the construction of an insulating surface lava layer (Figure 9a and 9b). The additional insulation provided by the lava layer will cause the ice-melting isotherm to ascend from its initial depth within the substrate (controlled by the surface temperature, overlying ice sheet thickness, geothermal heat flux, and thermal properties of the overlying materials) toward the base of the ice sheet, inducing bottom-up melting of the cryosphere and ice sheet (Figure 9a and 9b). The bottom-up melting rate is limited by the influx of geothermal heat, and therefore, it is possible for the insulating lava layer to be constructed more rapidly than the isotherm can rise, leading to “deferred” bottom-up melting (Figure 9c). If the insulating lava layer is constructed more slowly than the isotherm can rise, bottom-up melting will be insulation-limited (Figure 9d), and will proceed at a rate governed by lava accumulation.

The process of ice sheet lava heating and loading is ultimately controlled by the rate and volume of lava emplacement, factors that can vary greatly for different styles of volcanism. Consequently, in order to thoroughly assess the ice sheet lava heating and loading process, the nature of the volcanic processes generating lava emplacement must be constrained. The Hesperia Planum region (Figure 1) is the type location of the Hesperian ridged plains (Tanaka et al., 2014b), a volcanic plains unit characterized by the

presence of wrinkle ridge networks (*e.g.* Mangold et al., 1998), which is typical of Late Noachian-Early Hesperian volcanism, and is widely distributed across the surface of Mars (Head et al., 2002; Head et al., 2006c; Tanaka et al., 2014b). The Hesperian ridged plains are interpreted (Greeley and Spudis, 1981; Head et al., 2002) to have been emplaced in a flood basalt mode with voluminous eruptions of low viscosity magma at high effusion rates and with eruptive activity sustained over significant timescales. Measurements of linear ridges interpreted to be exposed dikes (Head et al., 2006c) associated with Early Hesperian plains in Tyrrhena Terra (an adjacent terrain on the western margin of Hesperia Planum) indicate the potential for lava effusion rates as high as $\sim 10^7$ m³/s (Wilson and Head, 1994). These effusion rate estimates are in general agreement with predictions derived from modeling (Keszthelyi et al., 2000; Keszthelyi et al., 2004) of martian flood basalt lava flows which suggest effusion rates on the order of 10^4 m³/s with surges reaching up to 10^5 to 10^6 m³/s. The interpreted mode of Hesperian ridged plains volcanism shares many similarities with the terrestrial large igneous provinces (which have been defined as flood basalt provinces covering an area greater than 10^5 km², with at least 1 km of accumulated lava cover, and with individual flows that exceed 100 km in length and 100 km³ in volume; Self et al., 1997), a relationship which has been previously recognized and explored (*e.g.* Head and Coffin, 1997; Head, 2006; Gregg, 2017) (Table 2).

While the general nature of Hesperian ridged plains volcanism is well established, precise measurements of many critical parameters are generally not available due to the preservation state of the plains units. These include: individual lava flow thicknesses, volumes, and areal extent, as well as the total timescale of volcanic activity, and the

frequency and duration of eruptive events. However, for these parameters, inferences can be drawn from the well-studied terrestrial large igneous province counterparts, and from measurements of the few examples of well-preserved flood basalt units (*i.e.* Elysium Planitia; Plescia, 1990; Keszthelyi et al., 2000; Sakimoto and Gregg, 2001; Berman and Hartmann, 2002; Burr et al., 2002a; Burr et al., 2002b; Fuller and Head, 2002; Plescia, 2003; Werner et al., 2003; Lanagan, 2004; Keszthelyi et al., 2004; Vaucher et al., 2009a; Vaucher et al., 2009b; Jaeger et al., 2010) and exposed dikes (*e.g.* Head et al., 2006c) on Mars.

The thicknesses of individual lava flows measured from terrestrial large igneous provinces range from ~1-150 m (Self et al., 1997; Sharma, 1997), with typical flow thicknesses in the Columbia River Flood Basalt province on the order of 10 m (Self et al., 1997). Note that for any given lava yield strength or surface slope, lava flows on Mars should be thicker (Wilson and Head, 1994) by a factor of ~2.5 due to the gravity scaling of lava rheology (Hulme, 1974). The areas and volumes of flow units measured from the Columbia River Flood Basalt province (Self et al., 1997) are on the order of 10^4 km², and 10^2 - 10^3 km³, which compare to the total (Coffin and Eldholm, 1994) province area of 1.6×10^5 km², and volume of 1.3×10^6 km³ (Table 2). The total area and volume of the Columbia River Flood Basalt province are relatively small on the scale of all terrestrial large igneous provinces (Table 2) which have total areas and volumes as large as 1.5 - 3.9×10^6 km² and 0.9 to $>2 \times 10^6$ km³ (Siberian Traps, the largest continental flood basalt province; Ross et al., 2005) to 4.88×10^6 km² and 36.4 - 76×10^6 km³ (Ontong Java, the largest oceanic flood basalt province; Coffin and Eldholm, 1994). Measurements of the Elysium Planitia flood basalt province on Mars (Plescia, 1990; Keszthelyi et al., 2000;

Sakimoto and Gregg, 2001; Berman and Hartmann, 2002; Burr et al., 2002a; Burr et al., 2002b; Fuller and Head, 2002; Plescia, 2003; Werner et al., 2003; Lanagan, 2004; Keszthelyi et al., 2004; Vaucher et al., 2009a; Vaucher et al., 2009b; Jaeger et al., 2010) indicate a total province area of $\sim 1\text{-}10 \times 10^6 \text{ km}^2$ and volume of $\sim 0.1\text{-}0.15 \times 10^6 \text{ km}^3$ (Table 2) with typical individual lava flow thicknesses of $\sim 1\text{-}40 \text{ m}$, areas of $\sim 10^4 \text{ km}^2$, and volumes of $\sim 10^3 \text{ km}^3$. The areas and volumes of the terrestrial large igneous provinces and the Elysium Planitia province are broadly similar to those of the Hesperia Planum ridged plains (Ivanov et al., 2005), which have a total area of $\sim 1.5\text{-}2 \times 10^6 \text{ km}^2$ and volume of $\sim 0.4\text{-}0.7 \times 10^6 \text{ km}^3$ (Table 2) (though the estimated volume of the Hesperia Planum ridged plains falls toward the low end of the terrestrial large igneous province range).

The total timescale of activity associated with the emplacement of terrestrial large igneous provinces is typically on the order of 10-100 Myr; however, the majority of eruptive activity and lava effusion generally occurs over relatively short intervals on the order of $\sim 1 \text{ Myr}$ (Coffin and Eldholm, 1994) (Table 2). Estimates from crater age dating (Vaucher et al., 2009a) place volcanic activity within the Elysium Planitia province as recently as $\sim 2 \text{ Ma}$ extending to at least 250 Ma. Individual periods of activity in the Elysium Province are estimated (Vaucher et al., 2009a) to be on the order of 1 Myr (Table 2), with some investigations (Jaeger et al., 2010) suggesting emplacement of large portions of the province (specifically the Athabasca Valles flood lava) over time periods as little as weeks to months. The timescale of Hesperian Ridged plains emplacement, and the plains of Hesperia Planum itself, are not thoroughly constrained. Previous estimates have suggested a total emplacement timescale of $\sim 100\text{-}200 \text{ Myr}$ (Craddock and Greeley, 2009; Tanaka et al., 2014a) for the Hesperian Ridged plains, with $\sim 0.01\%$ of that time

occupied by active eruptive phases, yielding a cumulative eruptive duration of 10 kyr (Halevy and Head, 2014). These estimates are supported by recent crater-size frequency analyses which indicate relatively uniform impact crater densities across the plains of Hesperia Planum, suggesting emplacement over a geologically short timescale (Mest et al., 2012; Gregg, 2017). The timescale estimates for the Hesperian Ridged plains are broadly consistent with both the total emplacement timescale and the proportion of active eruptions typical of the terrestrial large igneous provinces and the Elysium Planitia province. The estimated timescales of Hesperian Ridged plains emplacement, in conjunction with the ~1 km average total thickness of lavas within Hesperia Planum (Ivanov et al., 2005; Gregg, 2017), and assumptions of buildup of the unit by 10-100 m thick lava flows (consistent with measurements from the terrestrial large igneous provinces and Elysium Planitia), suggests a lava flow emplacement frequency of one flow per 100-1000 yr (for 10-100 m thick lava flows accreting to 1 km over a 10 kyr timescale) to one flow per 1-10 Myr (for 10-100 m thick lava flows accreting to 1 km over a 100 Myr timescale).

To assess ice sheet lava heating and loading as a formation mechanism for the upper Reull Vallis fluvial system both lava emplacement and ice sheet conditions are required. We first assume that ~0.3-1 km thick (Fastook and Head, 2015; Cassanelli and Head, 2016) regional Late Noachian ice sheets (predicted by climate modeling; Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015) covered the Hesperia Planum region prior to the emplacement of the Hesperian ridged plains. Glaciovolcanic interactions that could then occur during the subsequent emplacement of the effusive Hesperian ridged plain lavas depend upon the rate of lava emplacement, the thickness of

the lava flows, and the total accumulated thickness of lava (Cassanelli and Head, 2016). Measurements of buried and partially buried craters (Ivanov et al., 2005; Cassanelli and Head, 2016; Gregg, 2017) within Hesperian Planum indicate that the volcanic plains accumulated to a minimum average thickness of ~500 m (ranging from ~100-700 m) within the intercrater plains, and to ~2 km within crater interiors (ranging from ~0.5-4 km). There is no evidence to suggest that the source region of the upper Reull Vallis system is associated with an impact structure, therefore we assume the total accumulated lava thicknesses in this region are on the order of ~500 m. Under these assumptions four end-member ice sheet lava heating and loading scenarios are possible (Cassanelli and Head, 2016):

(1) Lava accumulated from thin (~10 m) flows relatively slowly (~100 Myr) (This would produce little top-down melting, instead causing steady bottom-up melting at relatively low rates throughout the emplacement period of 100 Myr).

(2) Lava accumulated from thin (~10 m) flows emplaced rapidly (~10 kyr) (This would again produce little top-down melting, instead causing steady bottom-up melting at relatively low rates throughout the emplacement period of 10 kyr and over a deferred melting period on the order of 10^6 yr).

(3) Lava accumulated from thick (~100 m) flows relatively slowly (~100 Myr) (This would result in rapid top-down melting following the emplacement of each flow until complete melting of the subjacent ice was reached).

(4) Lava accumulated from thick (~100 m) flows emplaced rapidly (~10 kyr) (This would again result in rapid top-down melting following the emplacement of each flow, but complete melting of the subjacent ice would occur more quickly given the shorter

total emplacement time).

Cassanelli and Head (2016) have previously explored the melting behaviors and morphologic consequences of ice sheet lava heating and loading under a variety of conditions including these end-member scenarios. Here, we are not concerned with evaluating the variability of melting behavior as a function of differing conditions, but instead aim to determine if ice sheet lava heating and loading is a viable mechanism to explain the formation of the Reull Vallis outflow channel system. Therefore, we focus specifically on these end-member cases and, following the justification outlined below, assess the most likely scenario to account for the formation of the Reull Vallis system.

We utilize the predictions of Cassanelli and Head (2016) to assess the consistency between the possible ice sheet lava heating and loading scenarios identified and the morphology of the upper Reull Vallis system. Despite variations in the lava emplacement conditions, each of the potential ice sheet lava heating and loading scenarios outlined above shares a number of predictions which are independent of the specific lava emplacement conditions. These mutual predictions are also generally consistent with several of the outlined upper Reull Vallis geomorphic constraints. Meltwater production from ice sheet melting by top-down contact heating will begin after individual lava flow emplacement events and, depending on lava flow thickness, may continue for up to several kyr (Cassanelli and Head, 2016). Meltwater production from ice sheet melting by bottom-up geothermal melting will begin after a sufficient thickness of lava has accumulated to raise the ice-melting isotherm to the base of the ice sheet which may take several 10^3 to 10^7 yr, with melting potentially delayed for an additional $\sim 10^3$ - 10^6 yr thereafter (Cassanelli and Head, 2016). In either case, ice sheet melting takes place after

lava flow emplacement (either individual flows for top-down melting, or thick sequences for bottom-up melting), and therefore the formation of the fluvial valleys will postdate lava flow emplacement for all ice sheet lava heating and loading scenarios, consistent with the first geomorphic criterion (the observed deformation of ridged plains lavas by the Reull Vallis fluvial system). In all scenarios, successive events of lava emplacement would result in successive episodes of heating and melting (Cassanelli and Head, 2016) and could account for the interpreted episodic nature of the upper Reull Vallis fluvial system, consistent with the second criterion (though the higher melt fluxes associated with top-down melting caused by the emplacement of thick lava flows would be more favorable for producing flooding conditions). Water generated by top-down and bottom-up heating and melting due to supraglacial emplacement of lava flows would result in subsurface flow (this would only occur for bottom-up melting if the substrate was relatively impermeable and inhibited infiltration) (Cassanelli and Head, 2016), thereby satisfying the third criterion.

While the preceding ice sheet lava heating and loading predictions are generally independent of the specific lava emplacement conditions, local and regional variations will result in differences in the predictions of meltwater generation and morphological outcomes. Differences in these predictions can affect the consistency between potential ice sheet lava heating and loading scenarios and the remaining upper Reull Vallis morphologic criteria (4 through 6). Ice sheet lava heating and loading is predicted to result in subsidence of the superposed lava flows due to melting of the subjacent ice sheet and removal of structural support (Cassanelli and Head, 2016). As a result, the fourth morphologic criterion, requiring subsidence within the northern trough source area, is

generally met by all ice sheet lava heating and loading scenarios. However, subsidence is only predicted to be significant during the accumulation of thin lava flows if a large total thickness is accumulated. This is because the accumulation of thin lava flows upon ice sheets acts to reduce direct top-down melting due to inefficient heat transfer (Cassanelli and Head, 2016), thereby preserving the ice beneath the lava flows. Therefore, if thin lava flows are accumulating, a large total thickness of lava must be accumulated in order to provide sufficient insulation to initiate bottom-up melting by geothermal heating to cause significant melting of subjacent ice sheets (Cassanelli and Head, 2016). Under typical predicted Late Noachian conditions (Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015), with ice sheets 300-1,000 m thick, this requires a minimum total thickness of ~0.6-1.1 km of lava (Cassanelli and Head, 2016) which exceeds the typical ~500 m thickness of the intercrater volcanic plains within Hesperia Planum (Ivanov et al., 2005; Cassanelli and Head, 2016; Gregg, 2017). Accumulation of this total thickness from relatively thin lava flows (~10 m or less in thickness) would result in a maximum of ~130 m of top-down melting and subsidence (Cassanelli and Head, 2016). Conversely, accumulation of the same 500 m total thickness from relatively thick (~100 m) lava flows would cause a maximum of ~1 km of melting and subsidence (Cassanelli and Head, 2016). Therefore, the ~100-200 m of subsidence observed within the northern trough is more consistent with an ice sheet lava heating and loading scenario involving the accumulation of thick lava flows which are able to cause more top-down melting and can more readily account for the observed subsidence.

Additionally, meltwater production through bottom-up basal melting is limited by the rate at which geothermal heat is input to the base of the ice, and therefore proceeds much

more slowly than top-down melting processes (Cassanelli and Head, 2016). Thus, providing meltwater fluxes sufficient to cause channel incision within Reull Vallis would require the confinement and storage of meltwater, followed by episodic releases. The martian regolith is predicted (Cassanelli and Head, 2016) to exhibit typical hydraulic conductivity values on the order of 0.01-10 mm/hr, far in excess of the melting rates generated by bottom-up basal melting which are estimated to be on the order of 0.01-1 mm/yr. Therefore, confinement of the meltwater produced by bottom-up melting requires the *ad hoc* presence of a relatively impermeable substrate to prevent the infiltration of meltwater, again favoring a top-down contact melting supply scenario.

The volume of the upper Reull Vallis fluvial system requires that a minimum of $\sim 2,300 \text{ km}^3$ of water was provided to erode the observed valleys. The estimated total lava thickness of $\sim 500 \text{ m}$ produces little insulation and, as a result, minimal bottom-up melting is predicted to occur, even if the regional ice sheets within the area had average thicknesses as great as 1 km (which would require highly asymmetrical ice accumulation or a very large surface water inventory; Fastook and Head, 2015; Carr and Head, 2015; Rosenberg and Head, 2015). Due to the limitations in melt volumes and melting rates, bottom-up basal melting is not predicted to be able to supply the water required for the formation of the upper Reull Vallis system. Consequently, top-down melting is favored as the primary source of meltwater generation to explain formation of the upper Reull Vallis system by ice sheet lava heating and loading.

Top-down melting under the Scenario (1) and (2) conditions involving the accumulation of thin lava flows is limited (because heat transfer from the heated flows is damped by previously accumulated and cooled lavas) and is therefore not favored to

account for the formation of the upper Reull Vallis system. In contrast, ice sheet lava heating and loading involving the accumulation of thick lava flows causes significant top-down contact melting as a result of direct heat transfer from the lava flows to the ice (Cassanelli and Head, 2016) (Figure 7), with greater sustained melting rates (Cassanelli and Head, 2016) (Figure 7). The cumulative volume of water which can be produced by ice sheet lava heating and loading is dependent upon the area over which melting occurs. The volume of the northern trough depression (Figure 4), the interpreted source of the upper Reull Vallis system, is estimated to be $\sim 426 \text{ km}^3$ (Kostama et al., 2007). In an ice sheet lava heating and loading scenario, this volume would represent the subsidence experienced by the volcanic plains due to the melting and removal of subsurface water ice within the northern trough source region. This volume is substantially less than the minimum $\sim 2,300 \text{ km}^3$ of water required to form the upper Reull Vallis fluvial system. Thus, in order for the upper Reull Vallis system to have formed through ice sheet lava heating and loading either, (1) the volume of the northern trough is not representative of the actual volume of subsurface water ice removed, or (2) ice sheet lava heating and loading occurred over a larger area.

To measure the area of the northern trough (evaluated here in the absence of previously reported measurements) we define the extent of the trough along the 1300 and 1250 m contours (Figure 9) which are equivalent to the elevation of the immediately adjacent wrinkle ridge plains. The defined extent of the northern trough gives an area of approximately $6,100 \text{ km}^2$ (Figure 10). Therefore, for melting confined within the bounds of the northern trough to produce the upper Reull Vallis system, an average of $\sim 410 \text{ m}$ of ice would need to be melted and removed within the northern trough area. However, this

would be likely to result in significantly more subsidence than the ~100-200 m observed within the northern trough. Conversely, if an average of 150 m of ice was removed from the subsurface, approximately equal to the average depth of the trough (Kostama et al., 2007), melting would have had to occur over an area of ~15,000 km², approximately 2.5 times larger than the observed northern trough area (context provided in Figure 11). If an average of 300 m of ice was melted and removed from the subsurface, a thickness equal to the approximate average predicted for Late Noachian regional ice sheets (Fastook and Head, 2015), melting over a ~7,500 km² area would be required, ~1.2 times larger than the northern trough area (context provided in Figure 11). Thus the volume of water required to form the Reull Vallis outflow channel system could not have been supplied by melting from ice sheet lava heating and loading exclusively within the Northern Trough.

Given the regional nature of the predicted Late Noachian ice sheets (Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015), and the Hesperia Planum ridged plains (Figure 1), lava-ice interaction and melting over a large area is plausible (Cassanelli and Head, 2016). In this case, ice sheet lava heating and loading could have been the source of the water required to form the upper Reull Vallis fluvial system through heating and melting over an area of ~7,500-15,000 km², with the northern trough acting as a conduit through which melt from this larger area was channeled to the surface to carve the upper Reull Vallis system. The northern trough, however, is clearly topographically depressed relative to the surrounding terrain. If melting occurred over the larger areas required to provide the minimum volume of water needed to carve the channel, this could be accounted for by the northern trough acting as a concentrated conduit of subsurface flow resulting in the scouring of additional subsurface material and

additional subsidence relative to the surrounding terrain. Subsidence and collapse features are not observed, however, over the areas of melting in the source region required to provide the volumes of water needed to carve the Upper Reull Vallis system, although the potential role of wrinkle-ridge formation that might accompany subsidence has not been evaluated.

Therefore, the most plausible scenario for the formation of the upper Reull Vallis by ice sheet lava heating and loading is one involving numerous phases of melting of thinner ice deposits which, after melting and removal, re-accumulate under an ambient “icy highlands” climate (Forget et al., 2013; Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth et al., 2015) and again interact with lava flows to generate continued melting. Concurrent ice and lava accumulation, the conditions which might promote it, and the effects it may have on ice sheet lava heating and loading are discussed in more detail by Cassanelli and Head (2016). In this scenario, reduced subsidence would be produced throughout the source region (due to the reduced thickness of the ice being melted) while the collapse associated with Waikato Vallis could have occurred from undercutting of surrounding ridged plains as the upper Reull Vallis channel underwent incision and lateral expansion.

We now explore the consistency of this most plausible scenario with the final criterion required by the geomorphology of the Upper Reull Vallis system, the fluid fluxes needed to sustain flow in the valley.

4.2 Ice Sheet Lava Heating and Loading: Melt Flux Model

To assess the meltwater volume flux that can be produced through ice sheet lava heating and loading we improve upon the melt flux estimates originally produced by

Cassanelli and Head (2016) through the implementation of a finite volume model to solve the 1-D transient heat conduction equation:

$$\rho(z) c_p(z) \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(K(z) \frac{\partial T}{\partial z} \right) \quad (2)$$

where ρ is density, z is depth, c_p is specific heat capacity, T is temperature, t is time, and K is thermal conductivity. The model is constructed with a 100 m thick layer of lava atop a large thickness of ice (to allow the ice to act as an infinite half-space) using a non-uniform grid spacing with spatial resolution decreasing with distance from the boundaries of the heated lava layer. The model grid construction is reviewed in more detail in a following section in which a model benchmarking analysis is performed. The thermal parameters adopted for the lava and ice in the model calculations are summarized in Table 3. The density, specific heat capacity, and thermal conductivity are held at constant values for the lava and ice layers individually (Table 3), but vary as a function of depth by changing from the lava values to the ice values at the lava-ice interface. An initial temperature of 1350 K is applied to the lava (Cassanelli and Head, 2016) while the ice is initialized with a steady-state temperature gradient increasing with depth from a surface temperature of 225 K (predicted Late Noachian mean annual surface temperature; Forget et al., 2013; Wordsworth et al., 2013; Wordsworth et al., 2015) with a geothermal heat flux of 55 mW/m² (approximate mean predicted Late Noachian value; Montési and Zuber, 2003; Ruiz et al., 2011). A constant temperature of 225 K is imposed at the surface of the lava layer for the upper boundary condition, while an insulating boundary condition is held at the base of the ice layer. The assumption of a constant surface temperature results in the overestimation of the early cooling rates of the lava flow.

However, modeling (Keszthelyi and Denlinger, 1996) and observations (Hon et al., 1994) have shown that initial lava flow surface temperatures rapidly decrease through radiative and convective heat loss, reaching near ambient values over timescales of days to weeks on the depth scale of the model grid cells we apply at the lava flow surface (0.02 m). This timescale is short relative to the ~ 100 yr total cooling timescale of the thick (100 m) lava flows we consider, and therefore should not have a strong influence on the overall cooling behavior. The phase change associated with the melting ice is accounted for by the heat integration method (*e.g.* Hu and Argyropoulos, 1996). The model does not account for the latent heat of fusion released by the cooling lava, therefore neglecting a portion of the heat energy that can be transferred from the lava to the ice. Assuming a typical basaltic composition with a liquidus of 1525 K, a solidus of 1225 K, a latent heat of fusion of 4×10^5 J/kg, and a linear release of latent heat between the liquidus and solidus, an initial lava flow temperature of 1350 K yields $\sim 1.6 \times 10^5$ J/kg of latent heat energy which is $\sim 16\%$ of the available sensible heat energy that can be transferred from the lava to the ice (given the temperature difference between the lava and ice, ~ 1125 K, and assuming a lava density of $3,000 \text{ kg/m}^3$ and specific heat capacity of 900 J/kg K). The latent heat released from the cooling lava would not increase peak melting rates but would result in a slower decay of the melting rates, leading to an increase in the average melting rate and a net increase in the total amount of potential ice melting.

Additionally, the model assumes that any ice which is completely melted is instantaneously removed from the system. Therefore, the model does not account for energy loss from the lava flow toward meltwater heating or steam creation, and thus the resulting melting rates and volumes reflect maximum end-member estimates. While this

simplifying assumption is made to facilitate the thermal modeling, and may be justified when lava-ice interactions occur over limited areas from which meltwater can readily escape, over the significant areas of lava-ice interaction we consider it will most likely be violated. This is evidenced by the presence of putative rootless cones in the Athabasca Valles lava flow (Lanagan et al., 2001; Keszthelyi et al., 2010) flood basalt unit of Elysium Planitia (generated through phreatomagmatic interactions of the Athabasca flood basalts and near-surface/surface water, most likely in the form of ice) which is similar in nature and scale to the Hesperia Planum ridged plains flood basalts we assess. The formation of rootless cones on Mars has been modeled by Dundas and Keszthelyi (2013) who found that the emplacement of lava flows could generate explosive pressures through the vaporization of subsurface ice beneath a dry porous regolith lag layer. However, at large lava flow thicknesses such as those considered here (~100 m), the generation of explosive pressures becomes more difficult, requiring the collection of steam from a broad area beyond the local eruption site or molten fuel-coolant interactions (caused by direct contact or intimate mixing of lava and ice) (Dundas and Keszthelyi, 2013). Molten fuel-coolant interactions are a clear consequence of the ice sheet lava heating and loading scenario we explore and therefore phreatovolcanic activity is reasonably expected, though the large thicknesses of the flood basalt lava flows could potentially suppress this activity. Despite this, there is no evidence in the vicinity of the Reull Vallis source region to suggest the occurrence of any phreatovolcanic processes, indicating that either they did not occur in this region, or if they did, the geomorphic evidence of such processes has been buried by the emplacement of subsequent lava flows. Therefore, while phreatovolcanic processes are a likely product of ice sheet lava

heating and loading which would consume energy from the lava flows, we do not consider them directly in our analysis and reemphasize that the melting rate and volume estimates we produce represent maximum values.

After initialization the model is run forward in time until melting is completed. Melting resulting from the emplacement of subsequent lava flows is modeled by adding a 100 m thick layer of cooled lava between the hot lava flow and the ice, which then serves as the surface from which the initial steady-state temperature gradient is applied, with all else proceeding as in the initial model. The melting rate and cumulative melt total estimates produced through this analysis for an emplacement sequence of three 100 m thick lava flows are shown in Figure 12.

4.3 Melt Flux Model Benchmark Analysis

In order to benchmark the finite volume numerical thermal model developed in this analysis, model results for a one-phase melting Stefan problem are compared against an analytical solution to assess consistency. The one-phase Stefan problem is solved analytically following the Neumann solution (*e.g.* Alexiades and Solomon, 1993) by introducing the similarity variable $\zeta = x / \sqrt{t}$, and seeking a solution for the melting-front position of the form $x(t) = A \sqrt{t}$ (where x is distance, t is time, and A is a constant). Setting: $A = 2 \lambda \sqrt{\alpha_L}$ (with λ a constant, and α_L the thermal diffusivity of the liquid) and defining the Stefan number: $St = c_L (T_L - T_m) / L$ (with c_L the specific heat capacity of the liquid, T_L an elevated boundary temperature, T_m the melting temperature of the solid, and L the latent heat of fusion) yields equations for the melting-front position x and for λ as follows (Alexiades and Solomon, 1993):

$$x(t) = 2 \lambda \sqrt{\alpha_L t} \quad (3)$$

$$\lambda e^{\lambda^2} \operatorname{erf}(\lambda) = \frac{St}{\sqrt{\pi}} \quad (4)$$

We select a benchmark problem for which the Stefan number is ~ 13.45 , which corresponds to a scenario involving the melting of a water-ice half-space with an initially uniform temperature distribution at the melting point to which a constant elevated temperature boundary condition is applied (with c_L the specific heat capacity of water, 4,186 J/kg K, T_L the elevated boundary temperature, set to 1350 K, T_m the melting temperature of water ice, 273 K, and L the latent heat of fusion of ice, 3.35×10^5 J/kg). This benchmark problem and the specific parameter values have been chosen to produce a scenario as close as possible to that assessed with the numerical model. The equation for λ is solved by the Newton-Raphson iterative method (using $\sqrt{St/2}$ as an initial approximation, which is the approximate solution for λ when the Stefan number is ~ 0 ; Alexiades and Solomon, 1993) and the computed value of λ is then used to solve equation (3) for the melting-front position as a function of time.

Numerical solutions to this one-phase Stefan problem have been generated using a range of non-uniform grid resolutions to assess solution convergence. The grids are constructed by setting an initial grid spacing at the elevated temperature boundary (Δx_o), and then adding cells with the spacing of each additional cell increased from the previous cell by a constant grid expansion ratio (GER) as expressed by the following relationship:

$$\Delta x_i = GER * \Delta x_{i-1} \quad (5)$$

The grids are constructed in this manner up to a total number of grid points (nx) required to reach the desired total thickness of the model domain. Numerical solutions have been computed for grids with Δx_o equal to 0.5, 0.1, and 0.02 m, with corresponding grid

expansion ratios of 1.075, 1.025, and 1.015, and nx of 60, 200, and 400 (Figure 13a). The time steps for each model have, respectively, been set to 0.5, 0.1 and 0.04 times the von Neumann stability criterion ($\Delta t \leq \Delta x^2 / 2 \alpha_L$), where the stability criterion is calculated using the Δx_o grid spacing (the minimum grid spacing in each model). Comparison of the numerical solutions shows convergence at increased spatial and temporal resolution and close agreement with the analytical solution for the benchmark problem (Figure 13b), supporting the validity of the numerical thermal model.

Given the accurate predictions of the high resolution numerical model grid on the benchmark problem, a similar grid has been constructed for application to the ice sheet lava heating and loading problem of interest (as described in Section 4.2). The grid is constructed by setting the initial grid spacing at the upper boundary of the heated lava layer to 0.02 m, then by adding additional grid cells with increased spacing using a grid expansion ratio of 1.015 as was done for the high resolution benchmark numerical model. Once the grid has reached a distance equal to half the thickness of the heated lava flow (50 m in this case), the grid expansion process is reversed and the grid spacing is progressively fined back to the initial grid spacing of 0.02 m at the bottom of the heated lava layer. The grid is then continued into the ice layer by setting the initial spacing of 0.02 m at the boundary and then proceeding with the grid expansion in the same manner until a desired ice thickness is reached (here 500 m, a thickness chosen to minimize the influence of the insulating bottom boundary condition without requiring a large number of additional grid cells). To model subsequent lava flows, the grid construction process remains identical, except that the grid expansion which initiated at the lava-ice boundary, now begins at the base of the heated lava layer at the boundary between the heated and

cooled lava. The grid configurations utilized in the modeling analysis are shown in Figure 13c.

Peak melting rates predicted by the thermal model (beyond the less accurate short timescales early in the model run), associated with the emplacement of the first lava flow, are approximately $5 \times 10^{-4} \text{ kg/m}^2\text{s}$, with average melt rates over the flow sequence of approximately $8 \times 10^{-6} \text{ kg/m}^2\text{s}$. In a scenario of lava accumulation and loading without contemporaneous ice accumulation, melting rates and volumes associated with subsequent flow emplacement are reduced as a result of inhibited heat transfer from the underlying cooled lava (Figure 12). In the favored scenario involving successive phases of ice accumulation and melting by lava flow emplacement, lava flows interact with the ice without considerable thicknesses of previously emplaced lava. Therefore, melting rates and volumes are represented by the model predictions for the first emplaced lava flow (Figure 12). Results are shown for 100 m thick lava flows but the same general patterns would be exhibited by thinner lava flows, with comparable melting rates but lower total melt volumes for any given flow due to less prolonged heat transfer (Cassanelli and Head, 2016).

4.4 Melt Flux Model Results

The melting rates predicted by the ice sheet lava heating and loading flux analysis (Figure 12) suggest that in order to supply the minimum estimated flux of the upper Reull Vallis system, melting would need to occur over areas of $\sim 10^7$ to 10^9 km^2 which are far in excess of the size of Hesperia Planum. The volume fluxes which can be produced by the predicted melting rates over the areas of required melting (the 6,100 km^2 northern trough, and the 7,500 km^2 and 15,000 km^2 areas; Figure 11) are summarized in Table 4. The

volume flux estimates are shown along with the corresponding water depths that could be sustained by each flux within the minimum channel dimensions measured from the upper Reull Vallis system (as discussed in Section 3). While the volume flux of the upper Reull Vallis system estimated through the application of the Darcy-Weisbach equation assuming bankfull flow is clearly an overestimate, the meltwater fluxes provided by the ice sheet lava heating and loading mechanism are several orders of magnitude below this estimate. As such, unless the Reull Vallis system could have been carved with greatly reduced water fluxes with water depths on the order of 1 m (assuming full channel width), ice sheet lava heating and loading would only have been able to supply sufficient meltwater fluxes if the meltwater underwent episodes of confinement and release.

In general, we find that an ice sheet lava heating and loading origin for the upper Reull Vallis offers advantages over previous formation models (*e.g.* Squyres et al., 1987), primarily in the ability to supply the requisite volumes of water. However, the apparent lack of extensive subsidence/collapse in the channel source region (not accounting for the possible contribution from wrinkle ridge formation) and relatively low meltwater fluxes that are generated suggest that, while this may be an important contributing mechanism, other processes were likely to have been involved in the formation of the upper Reull Vallis system. What other processes may have been involved, and could they augment the ice sheet lava heating and loading mechanism to explain the formation of the outflow channel system? To address this question we now explore flux modulation by meltwater impoundment and release as well as the formation of the Reull Vallis outflow channel by a series of many lower-discharge flows (a mechanism that has been suggested by modeling of the circum-Chryse outflow channels; Andrews-Hanna and Phillips, 2007;

Harrison and Grimm, 2008) as plausible additional processes that may aid in explaining the origin of the upper Reull Vallis.

5. Flux Modulation by Impoundment and Release

Results of the thermal modeling analysis indicate that the melt fluxes generated by lava ice heating and loading may be insufficient to solely account for the formation of the Reull Vallis outflow channel without the operation of flux modulation processes. The discharge rates produced by ice sheet lava heating and loading could have been increased by confinement and build-up of meltwater with subsequent release, as has been invoked to explain the formation of other outflow channels (Robinson and Tanaka, 1990; Burr et al., 2002a; Harrison and Grimm, 2008; Coleman and Baker, 2009; Marra et al., 2014). We now review potential mechanisms that could have caused local meltwater confinement and release during ice sheet lava heating and loading and the formation of the upper Reull Vallis system.

(1) Meltwater confinement and accumulation could have occurred within supraglacial or proglacial surface reservoirs created by snow, ice, or sediment/regolith damming with subsequent failure generating significantly increased meltwater discharge:

In this wide range of possible confinement scenarios, failure could have occurred by (a) disruption of dams by melting (*e.g.* Benito and O'Connor, 2003; O'Connor and Costa, 2004), (b) increased water volumes and pressure (*e.g.* Baker and Milton, 1974; O'Connor and Costa, 2004), (c) drainage through ice sheet fractures, moulins, or permeable layers (*e.g.* Smellie, 2001; Das et al., 2008; Liang et al., 2012; Doyle et al., 2013; Tedesco et al., 2013), (d) triggering from external mechanisms (Harrison and Grimm, 2008) such as landslides, avalanches, or ice mass failures delivering additional mass into the reservoir,

or (e) increased pore pressures within the dam materials (Costa and Schuster, 1988). Meltwater confinement and release in this range of scenarios is thought to have been responsible for the formation of the terrestrial Channeled Scablands of eastern Washington State, a terrestrial analog to the martian outflow channels produced by disruption and catastrophic flooding from Pleistocene Lake Missoula (Bretz et al., 1956; Baker and Milton, 1974; Baker, 1973; Baker and Bunker, 1985; Burr et al., 2009).

(2) Meltwater confinement and accumulation could have occurred within englacial or subglacial reservoirs, with subsequent reservoir failure generating meltwater discharge:

Possible englacial or subglacial reservoirs include (a) permeable snow, firn, and ice layers (*e.g.* Forster et al., 2014; Cassanelli and Head, 2015), (b) fractures, moulins, and crevasses (Cuffey and Paterson, 2010), and (c) subglacial voids, lakes, and drainage systems (Cuffey and Paterson, 2010). Failure of these reservoirs could have occurred by drainage through ice sheet fractures, moulins, permeable layers, or subglacial drainage networks (*e.g.* Smellie, 2001; Smellie, 2006; Das et al., 2008; Liang et al., 2012; Doyle et al., 2013; Tedesco et al., 2013) and by the pressurization of subsurface reservoirs through meltwater addition causing floatation of the confining ice and rapid subglacial drainage, a primary mechanism for the generation of jökulhlaups (large glacial outburst floods on Earth and Mars) (*e.g.* Björnsson, 1988; Björnsson, 2003; Smellie and Chapman, 2002; Smellie, 2009; Scanlon et al., 2014; Smellie and Edwards, 2016).

(3) Subglacial and englacial magmatic intrusion and melting: This mechanism can lead to the production of subglacial and englacial meltwater pools (Wilson and Head, 2002) which can undergo failure and release by the same processes as previously discussed (with the addition of the potential for volcanic melting to extend toward the ice

cap margin, opening a pathway to the surface and discharging stored meltwater) (Wilson and Head, 2002). Melting produced by subglacial and englacial volcanism is predicted to be more efficient than melting induced by top-down heating from lava loading, as a greater proportion of the heat contained within the lava is transferred to the ice (Squyres et al., 1987). As a result, subglacial and englacial volcanism and ice melting can produce larger volumes of meltwater and greater discharge rates than top-down lava loading and heating for any given melting area or volume of lava/magma (though the total melt volumes produced would be limited by the size of the intrusive body).

The discharge rates produced by these confinement and release mechanisms depend exponentially upon the size of the breach through which water is released after storage (O'Connor and Costa, 2004; O'Connor and Beebee, 2009). Discharge can reach rates on the order of $10^7 \text{ m}^3/\text{s}$, and is generally highest for surface reservoir failure and jökulhlaup-type subglacial outbursts (O'Connor and Costa, 2004). Therefore, it is possible for discharge rates on the order of the peak discharge estimates measured from the upper Reull Vallis outflow channel system to have been achieved by these meltwater confinement and release mechanisms. In addition, melting over short timescales, as is predicted to occur from lava-ice interactions, does not require confining geometries to store water for great lengths of time to accumulate large water reservoirs. Therefore, there is strong potential for confining reservoir formation and failure to result from top-down ice sheet melting induced from lava emplacement and heating. However, while there is potential for meltwater impoundment and release to increase the discharge rates supplied by the ice sheet lava heating and loading mechanism, there is no direct geomorphic evidence to indicate that this process has occurred. We now proceed to a

final assessment investigating the potential formation of the Reull Vallis by multiple flooding events as a process that may aid in achieving the necessary discharge requirements of the outflow channel system.

6. Reull Vallis Formation by Multiple Floods

In order to explain the formation of the Reull Vallis system by a series of many lower-discharge flows, the individual lower discharge flows must attain some minimum discharge threshold to produce the observed erosion. This minimum discharge threshold can be estimated as the minimum discharge required to produce the necessary bed stress to begin sediment motion, an often utilized method (Baker, 1973; Komar, 1979; Costa, 1983; O'Connor, 1993; Carling, 1996; Lamb et al., 2008; Lamb and Fonstad, 2010). Here we calculate the minimum discharge in this manner following Lamb et al., (2008) and Lamb and Fonstad (2010). Assuming hydraulically rough, steady and uniform flow, discharge is estimated as (Parker, 1991):

$$Q = 8.1 A \left(\frac{\tau_b}{\rho_f} \right)^{1/2} \left(\frac{d}{k_s} \right)^{1/6} \quad (6)$$

where A is the cross-sectional area of the flow (m^2), τ_b is the bed-shear stress (N/m^2), ρ_f is the fluid density (here assuming $1,000 \text{ kg/m}^3$), and k_s is the roughness-length scale of the channel bed (m). Incipient sediment motion occurs when the Shields parameter (τ_*) exceeds a critical value (τ_{*c}) which is calculated as a function of the bed-slope gradient (S) following Lamb et al. (2008b), with the critical value given by:

$$\tau_{*c} = 0.15 S^{0.25} \quad (7)$$

The bed shear stress corresponding to the critical Shields value is then calculated with the

following relation (Lamb et al., 2008a):

$$\tau_{*c} = \tau_b / ((\rho_s - \rho_f) g D_{50}) \quad (8)$$

where ρ_s is the sediment density (kg/m^3) and D_{50} is the median grain diameter (m). The bed shear stress required for sediment motion is then used to calculate the hydraulic radius from (Lamb et al., 2008a):

$$\tau_b = \rho_f g h_R S \quad (9)$$

The computed hydraulic radius is then used to compute the flow depth, and finally, the minimum discharge required for sediment motion is determined by equation (5). For the Reull Vallis outflow channel, we find a minimum flow depth of ~ 1.5 m, and a minimum discharge of $\sim 2.2 \times 10^4 \text{ m}^3/\text{s}$ (assuming a rectangular channel cross-section) with $W=10,800$ m, $S=0.15^\circ$ (W and S measurements taken from location 2; Figures 3-5, Table 1), $(\rho_s - \rho_f)=1,800 \text{ kg/m}^3$ for basalt (Lamb et al., 2008a), $D_{50}=0.064$ m (the average D_{50} value measured from the Viking 1, Viking 2, and Pathfinder landing sites; Wilson et al., 2004), and $k_s=0.164$ m (assuming $k_s \sim D_{84}$ and using the average D_{84} value measured from the Viking 1, Viking 2, and Pathfinder landing sites; Bathurst, 2002, Wilson et al., 2004). Minimum discharge estimates have been computed for the location 2 site instead of location 3 for simplicity in performing the assessment with one channel and because of the comparable peak fluxes estimated at location 2 and 3 (Table 1).

The derived minimum discharge estimate for incipient sediment motion is approximately a factor of 3 greater than the maximum fluxes predicted to have been

generated by ice sheet lava heating and loading within the Reull Vallis source region (Table 4). This minimum discharge estimation relies on constraints of the characteristic size of particles moved by the flow as well as the channel width, both of which are not well defined for the period of formation of the Reull Vallis system (as the characteristic grain sizes eroded by the flow are unknown, and because the channel width evolved throughout formation and has likely undergone post-flow erosion and modification). Repeating the minimum discharge calculation for a median grain diameter of 10 cm (a value that has been previously assumed for martian fluvial channels; Kleinhans, 2005) yields a minimum flow depth of ~ 2.3 m and a discharge of $\sim 4.7 \times 10^4$ m³/s, and for a channel width of 5,400 m (approximately half the measured width of Reull Vallis at location 2, accounting for the possibility of significant post-flow erosion and modification) yields a minimum flow depth of ~ 1.5 m and discharge of $\sim 1.1 \times 10^4$ m³/s.

To assess a broader potential range in minimum discharge estimates derived from the incipient sediment motion threshold, a sensitivity analysis accounting for a conservative measurement error and an expanded parameter space has been performed. Minimum discharge estimates have been calculated (following the procedure outlined by equations 6-9) for the upper Reull Vallis system at location 2 for a range of plausible channel slope values allowing for a width measurement error fraction of ± 0.2 and an order of magnitude variation in the D_{50}/D_{84} bed grain size parameters (Figure 14). The evaluated variation in the D_{50} and D_{84} bed grain size parameters yields ranges of 0.0064-0.64 m for the D_{50} grain size and 0.0025-0.25 m for the D_{84} grain size. Similarly, an order of magnitude variation in the channel slope from the location 2 measurement (0.15° ; Figure 14a) has been assessed, with a minimum slope of 0.015° (Figure 14b) and a maximum

slope of 1.5° (Figure 14c). The sensitivity analysis (Figure 14) illustrates the dependence of the minimum discharge estimate upon channel width, slope, and the D_{50} and D_{84} bed grain size parameters, with the strongest control exerted by the grain size parameters. The minimum discharge estimate increases with increasing D_{50} , D_{84} , channel width, and decreasing channel slope.

The sensitivity analysis (Figure 14) indicates that for the range of measurement error considered, and the plausible range of input parameters, the incipient bed motion minimum discharge estimates cover a wide range from $\sim 10^2$ to 10^6 m^3/s . The calculated range of minimum discharge estimates spans two orders of magnitude above and below the nominal minimum discharge estimate (of order 10^4 m^3/s), resulting largely from variations in the poorly constrained D_{50} parameter. These minimum discharge estimates could plausibly be supplied by ice sheet lava heating and loading in the Reull Vallis source region up to the nominal estimate of $\sim 10^4$ m^3/s (Table 4), but greater discharge values are prohibitively large, approaching the peak discharge values estimated earlier under the bankfull channel assumption. However, the estimated minimum discharge values only exceed the nominal value at relatively large median bed grain sizes on the order of 10 cm. Therefore, for a reasonable range of median grain diameters less than ~ 10 cm, these results suggest that the discharge provided by ice sheet lava heating and loading in the Reull Vallis source region could plausibly have produced the erosion observed in the outflow channel system. Formation of Reull Vallis at these lower discharges would require a greater number of flow events and more sustained flow activity in order to provide the necessary cumulative volumes of water to erode the valley. As discussed earlier (in Section 4.1), multiple discharge events could have been

produced by the successive emplacement of Hesperian ridged plains lava flows. However, depending upon the number of flooding events required to produce the outflow channel system, this could require the emplacement of thinner individual flows (which produce comparable melting rates but less overall melting for each individual lava flow event; Cassanelli and Head, 2016) to increase the number of discharge events produced during the accumulation of the 500 m thick lava unit. The requirement of thinner individual lava flows would in turn require concurrent lava and ice accumulation (Cassanelli and Head, 2016) so that the accumulation of previously-emplaced cooled lava would not inhibit top-down heat transfer and melting of the ice (as discussed in Section 4.1). Additionally, in this scenario each individual lava flow would only experience a limited amount of subsidence and would not be likely to produce the extensive collapsed geomorphology observed in Waikato Vallis. Therefore, subsidence and collapse would need to be generated by some other mechanism, with a likely candidate being the undercutting of surrounding ridged plains during incision and lateral expansion of the upper Reull Vallis channel.

7. Conclusions

Geomorphic assessment of the upper Reull Vallis outflow channel system has been performed to identify several unique criteria that must be met by plausible formation mechanisms: (1) Formation post-dating Hesperian ridged plains emplacement, as evidenced by the deformation and erosion of the wrinkle ridge plains near the channel head, and the stratigraphic superposition of the channel downstream. (2) Episodic release of water, consistent with observations suggesting multiple periods of flooding activity, suggested by the presence of terraces within Waikato Vallis and Segment 2. (3)

Generation of subsurface flow and undercutting as indicated by the subsidence and collapse features in the northern trough and Waikato Vallis. (4) Subsidence of $\sim 100\text{--}200$ m within the northern trough channel head, with retention of the Hesperian ridged plains surface morphology evidenced by the morphology of the northern trough. (5) Production and release of a minimum of $\sim 2,300 \text{ km}^3$ of water, an amount equal to the minimum amount of water needed to erode the volumes of the upper Reull Vallis channels. (6) Supply of a minimum discharge of $\sim 10^4 \text{ m}^3/\text{s}$, indicated by the minimum flux required to initiate channel bed sediment motion, and a maximum discharge of $\sim 10^7 \text{ m}^3/\text{s}$, indicated by flux estimates derived from the most conservative measurements available for the upper Reull Vallis system.

Previous models for outflow channel formation in the Hesperia Planum region, which have invoked melting of ground-ice through intrusive magmatism as a source mechanism, do not adequately account for the required volumes of water and the fluvial morphology of the outflow channel system. We explored the potential genetic relationship between Hesperia Planum volcanism and the upper Reull Vallis fluvial system and assessed an ice sheet lava heating and loading model as a potential alternative formation mechanism. We find that a scenario involving sequential phases of melting ice deposits, which re-accumulate under an ambient “icy highlands” climate (Forget et al., 2013; Wordsworth et al., 2013; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth et al., 2015) to undergo continual lava-ice interactions is most consistent with the geomorphology of the upper Reull Vallis system. This ice sheet lava heating and loading mechanism offers several advantages over previous formation models, primarily in terms of its ability to supply the requisite volumes of water (since the ice volume is not

constrained by the subsurface pore space) and its broadly consistent morphologic predictions (subsidence and collapse predicted in source region as well as more appropriate water-to-sediment ratios). However, the apparent lack of sufficient subsidence/collapse in the channel source region, the relatively low meltwater fluxes generated, and the possibility of water storage areas provided by this mechanism represent areas for future analysis in order to understand the complete range of mechanisms responsible for the formation of the upper Reull Vallis system.

Acknowledgements

We are grateful to Lionel Wilson and Laszlo Kestay for thorough and constructive reviews which improved the quality of the manuscript. We thank David Weiss, James Russell, and Marc Parmentier for helpful comments and discussion. We acknowledge support to JWH for participation in the Mars Express High-Resolution Stereo Camera Team (JPL 1237163).

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Waikato Vallis Measurements

Location	Depth (m)	Width (m)	Slope (°)	Flow Velocity (m/s)	Volume Flux (m ³ /s)
1	120	11,800	0.15	21.3	3.0x10 ⁷
2	80	10,800	0.15	16.9	1.5x10 ⁷
3a	25	4,500	0.40	12.7	1.4x10 ⁶
3b	100	2,900	0.40	30.1	8.7x10 ⁶
3 Total					1.0x10 ⁷

Segment 2 Measurements

Location	Depth (m)	Width (m)	Slope (°)	Flow Velocity (m/s)	Volume Flux (m ³ /s)
1	160	8,200	0.35	38.5	5.1x10 ⁷
2	200	5,100	0.22	34.2	3.5x10 ⁷
3	290	8,600	0.25	46.2	1.1x10 ⁸

Table 1. Valley dimension measurements for the upper Reull Vallis outflow channel system (measurement locations marked in Figures 3-5) and flux estimates derived from the Darcy-Weisbach equation assuming bankfull flow. Note that measurements from locations 3a and 3b (shown in Figures 4 and 5) reflect a branching point of Waikato Vallis and are therefore summed to produce a total flux estimate. Additionally, location 3 is a point at which channel terraces are most apparent within the upper Reull Vallis system (Figures 4 and 5). Therefore, measurements at this location were taken from the innermost terraces in an attempt to define a minimum flux estimate.

Province	Area (10 ⁶ km ²)	Volume (10 ⁶ km ³)	Age Range (Ma)
<i>Terrestrial Provinces, Continental</i>			
Columbia River Flood Basalts ^a	0.16	1.3	6-17.5; 15.7-17.2
Deccan Traps ^a	0.8-1.8	8.2	64.5-65.5; 65-69
Siberian Traps ^{b,c}	1.5-3.9	0.9 to >2	249-250*
<i>Terrestrial Provinces, Marine</i>			
Kerguelen ^a	2.3	15-24	109.5-114
Ontong Java ^a	4.9	36-76	117.7-118.2; 121-124
<i>Martian Provinces</i>			
Elysium Planitia ^d	~1-10	~0.1-0.15	~2-250
Hesperia Planum ^{e,f}	~1.5-2	~0.4-0.7	?

^aCoffin and Eldholm (1994), ^bRoss et al., (2005), ^cSharma (1997), ^dReferences provided in Section 4.1, ^eIvanov et al., (2005), ^fGregg (2017)

Table 2. Areas, volumes, and age ranges for a number of terrestrial large igneous provinces and for comparable martian volcanic provinces. *The indicated age range for the Siberian Traps is based on an established inception age of 250 Ma and an emplacement timescale of 1 Myr for the bulk of the eruption (Sharma, 1997).

Parameter	Lava	Ice
<i>Thermal Conductivity (W/m K)</i>	2	2
<i>Density (kg/m³)</i>	3,000	917
<i>Specific Heat Capacity (J/kg K)</i>	900	2,000
<i>Latent Heat of Fusion (J/kg)</i>	--	3.35x10 ⁵

Table 3. Thermal parameters used in the ice sheet lava heating and loading melt flux model.

Source Area	Area (m ²)	Max flux (m ³ /s)	Mean flux (m ³ /s)	Max fill depth (m)	Mean fill depth (m)
<i>Northern Trough</i>	6.1x10 ⁹	3,300	50	0.6	<0.1
<i>7,500 km²</i>	7.5x10 ⁹	4,100	65	0.7	0.1
<i>15,000km²</i>	1.5x10 ¹⁰	8,200	130	0.9	0.2

Table 4. Meltwater volume flux estimates generated from maximum and mean melting rate predictions from the ice sheet lava heating and loading melt flux model applied over the estimated source areas of the upper Reull Vallis system. Also shown are the approximate water depths which could be sustained by these volume fluxes within the minimum channel dimensions (Table 1 and Figure 5) established for upper Reull Vallis (measured from location 3; Figures 3-5).

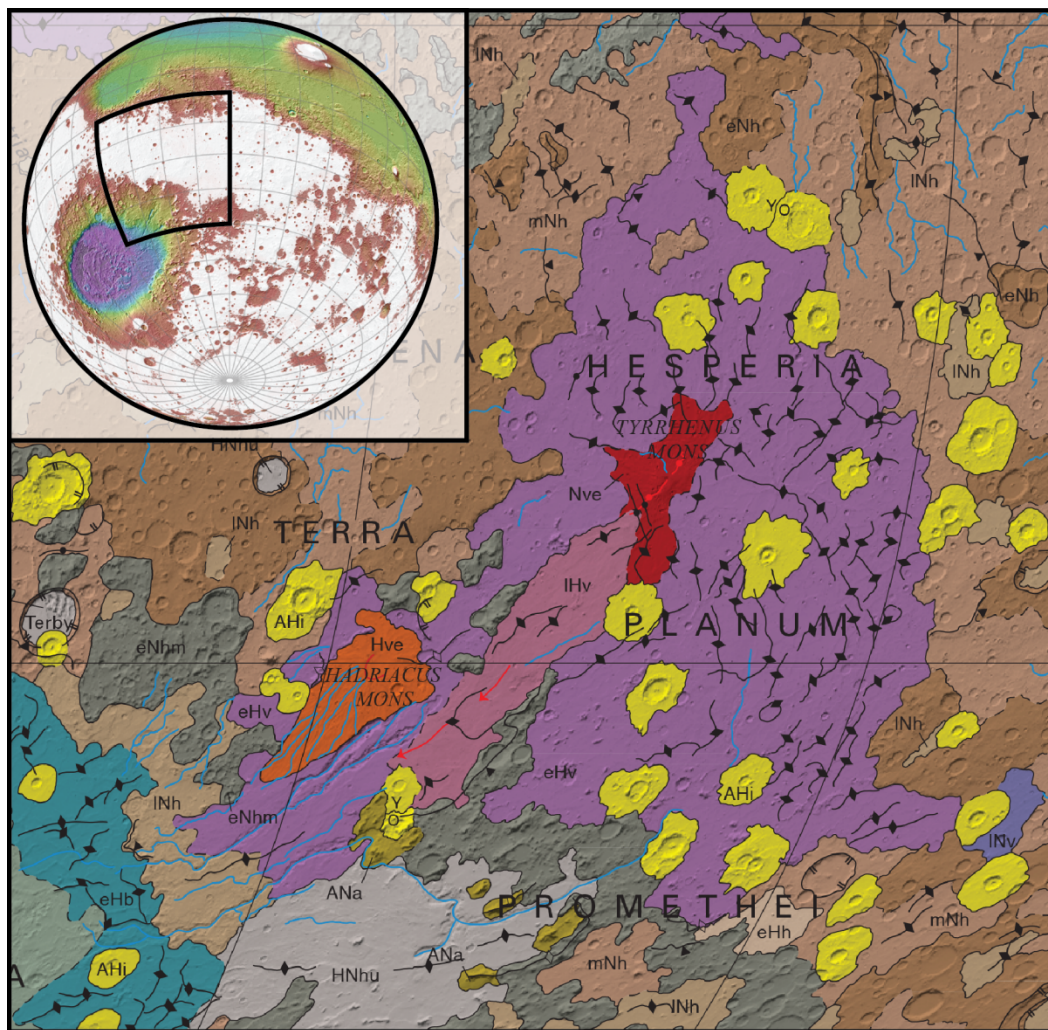


Figure 1. Geologic map of the Hesperia Planum region (Tanaka et al., 2014b) with inset image for global context. Inset shows a topographic map of Mars depicting a conceptual representation of the “icy highlands” (Head and Marchant, 2014) early climate scenario (Forget et al., 2013; Wordsworth et al., 2013; Fastook and Head, 2015). All regions above the predicted +1 km equilibrium line altitude (ELA), above which snow and ice are predicted to be characterized by annual net accumulation, are shaded white.

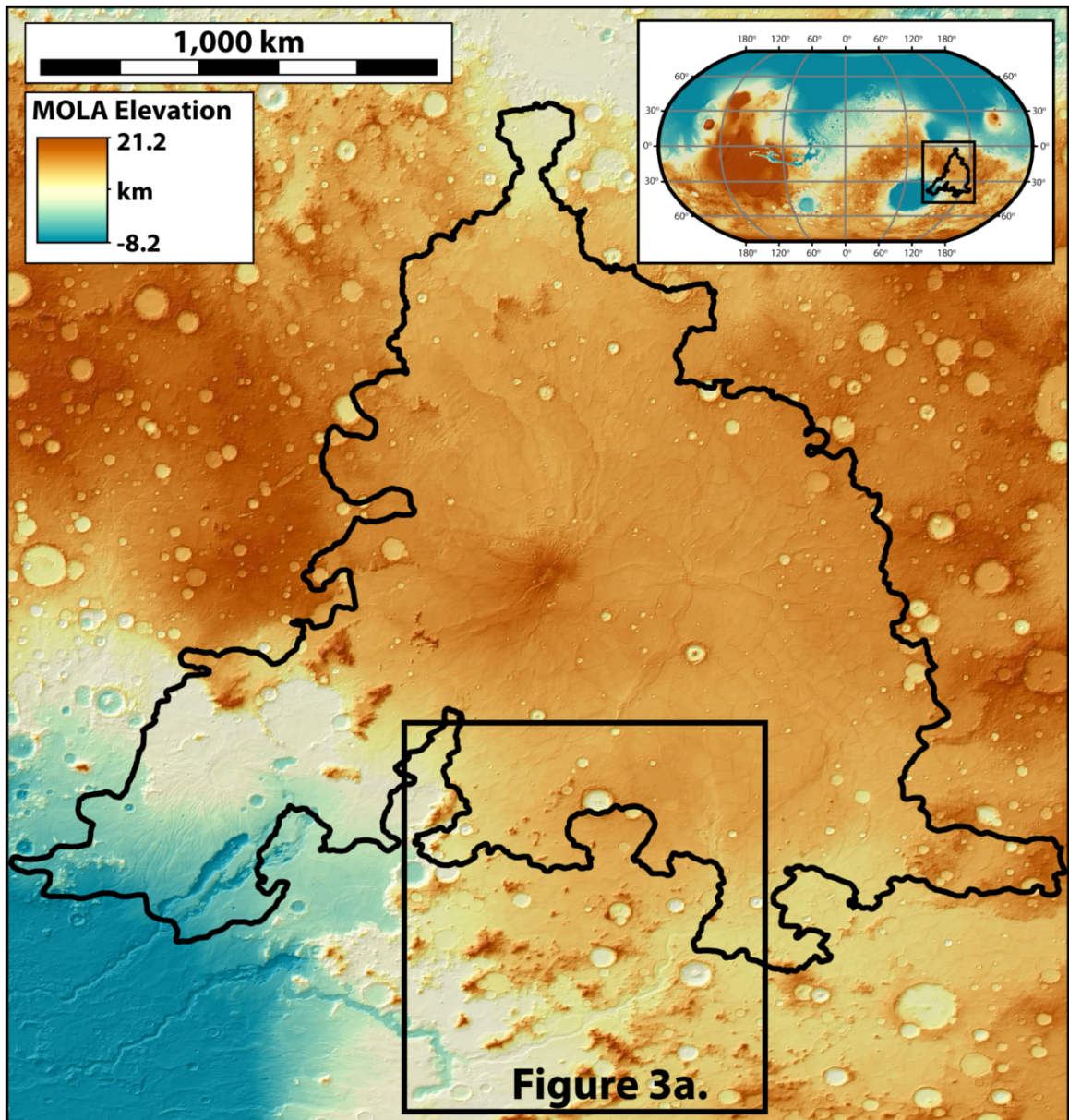


Figure 2. The outlined extent of the Hesperia Planum region (Tanaka et al., 2014b) shown in MOLA hillshade overlain by gridded MOLA topography (Smith et al., 2001). Global context of the Hesperia Planum region is provided in the inset panel showing a shaded MOLA relief map of Mars. The extent of Figure 3a is delineated to provide contextual reference.

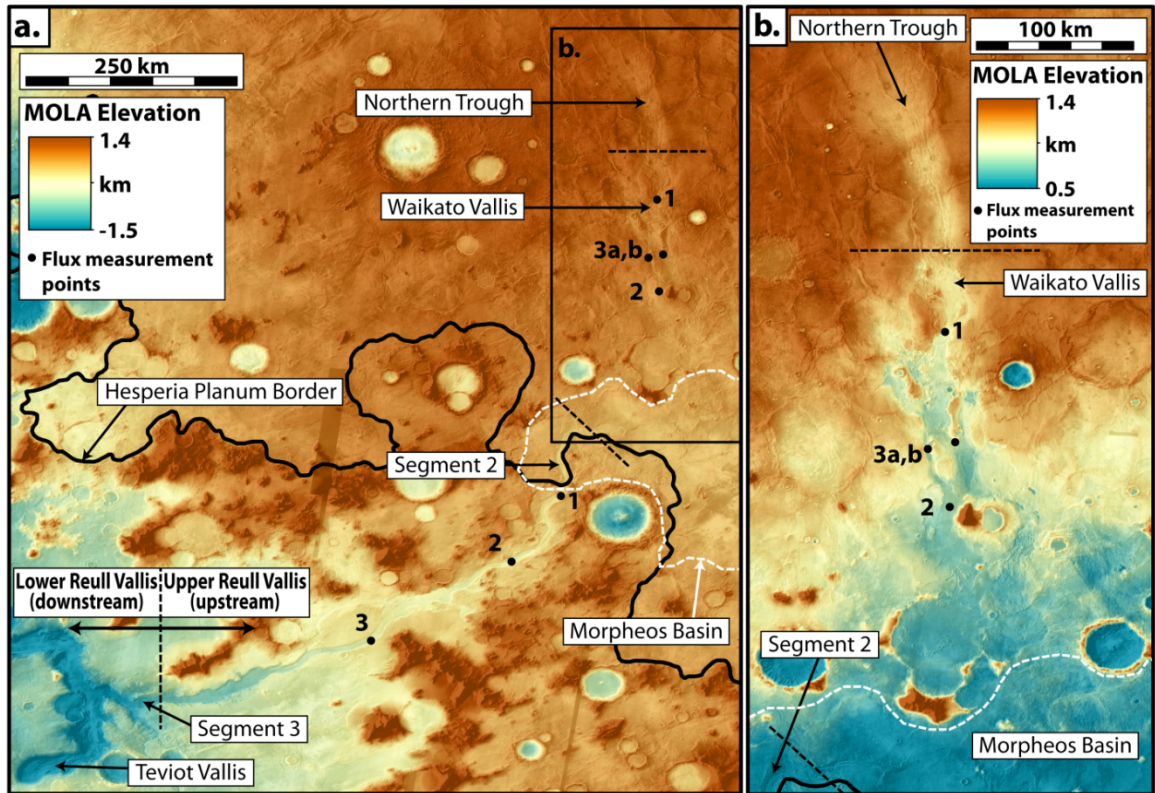


Figure 3. (a.) The Reull Vallis fluvial system in the eastern rim region of the Hellas basin (context provided in Figure 2) shown in a THEMIS global daytime infrared mosaic (Christensen et al., 2004) with overlain gridded MOLA topography data (Smith et al., 2001) (used here and in subsequent figures unless otherwise noted). The individual segments which comprise the ~1,700 km long Reull Vallis fluvial system (Kostama et al., 2007; Kostama et al., 2013), divided into upper and lower portions, are labeled. Waikato Vallis is formerly known as Segment 1 (Kostama et al., 2007). **(b.)** Inset from panel (a.) providing a more detailed view of the upper portion of the Reull Vallis fluvial system. Locations from which valley dimension measurements have been taken to produce flux estimates are indicated.

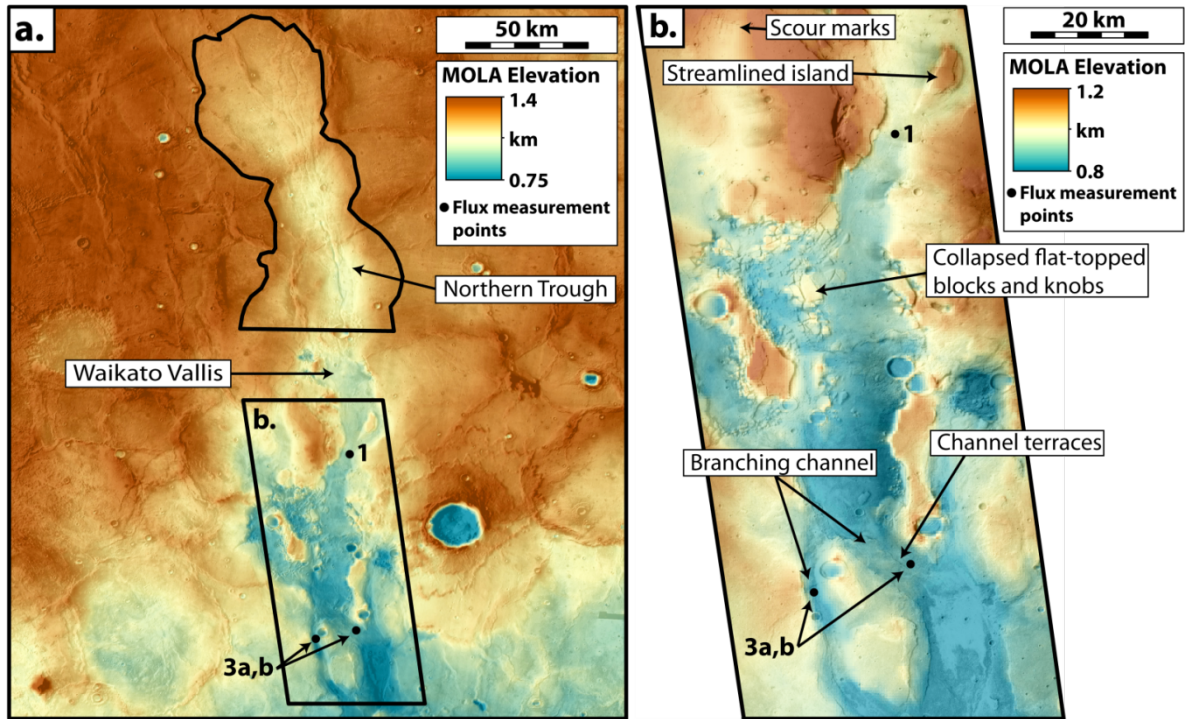


Figure 4. (a.) The head of the upper Reull Vallis fluvial system within Hesperia Planum. The northern trough has been interpreted as the source for the water which carved the upper Reull Vallis system. The upper Reull Vallis system emerges abruptly as a fully-sized fluvial valley within Waikato Vallis (formerly Segment 1 of the system). The extent of the northern trough is delineated along the 1250 and 1300 m contours, covering an area of $\sim 6,100 \text{ km}^2$ (see Figure 10). (b.) Inset panel from (a.) showing a portion of Waikato Vallis in a CTX (Context Camera; Malin et al., 2007) mosaic (CTX images D12_032015_1470_XI_33S246W and D12_031949_1484_XI_31S246W) with overlain gridded MOLA topography data. The fluvial and collapse/subsidence-related features discussed in Section 3 are indicated.

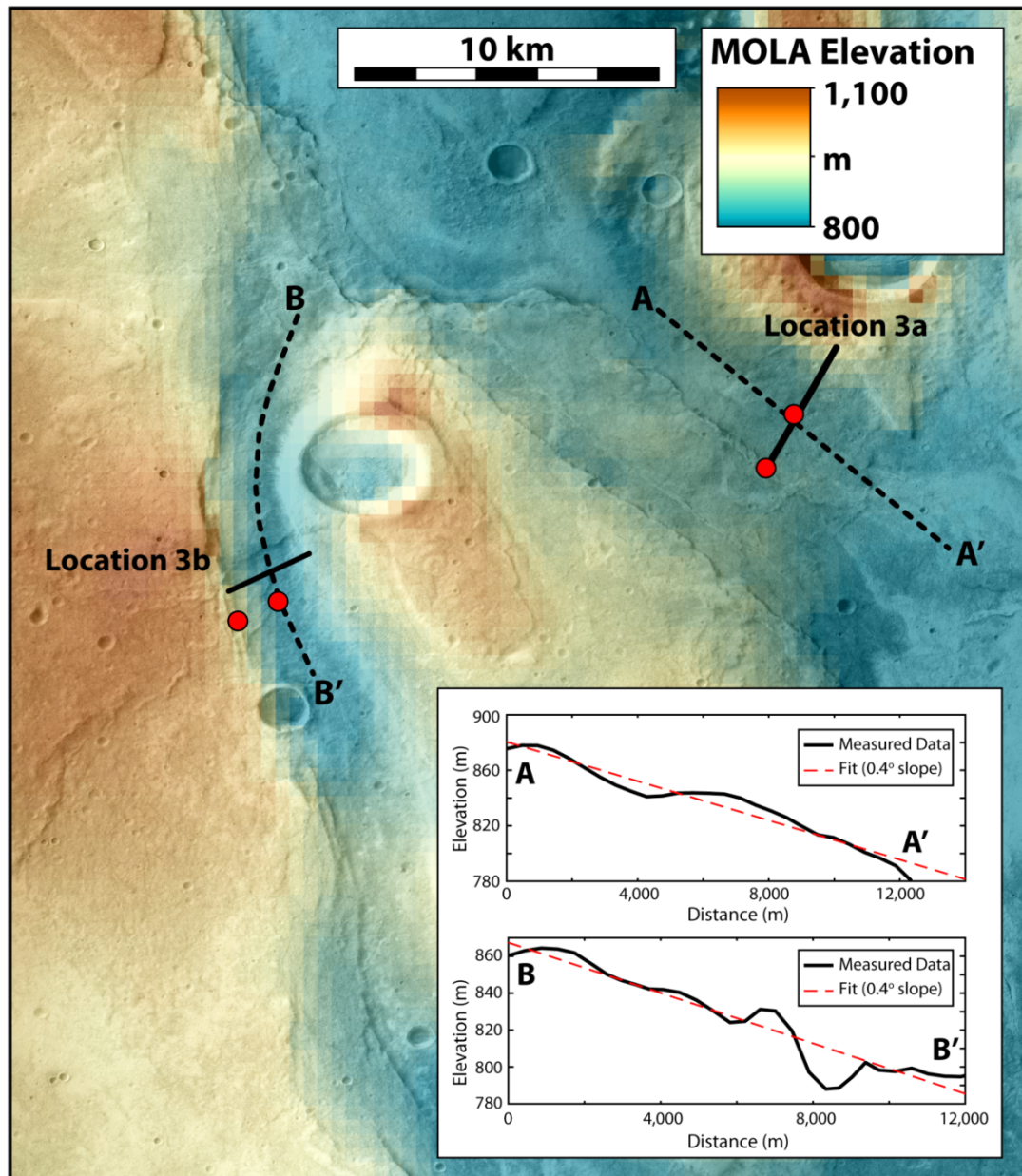


Figure 5. Example illustrating the channel flux measurement procedure for the upper Reull Vallis outflow channel system. Red dots indicate the individual MOLA shot points used to measure the depth of the channel, bold black lines indicate the width measurements taken from the CTX images (CTX images D12_032015_1470_XI_33S246W and D12_031949_1484_XI_31S246W), and the dashed black lines show the gridded MOLA profiles taken to estimate the channel slope (A-A' and B-B'). Valley elevation profiles A-A' and B-B', shown in the insets, are fit with a linear trend (dashed red line) to derive the valley slope estimates. All data and measurements that were collected to estimate the channel flux are reported in Table 1.

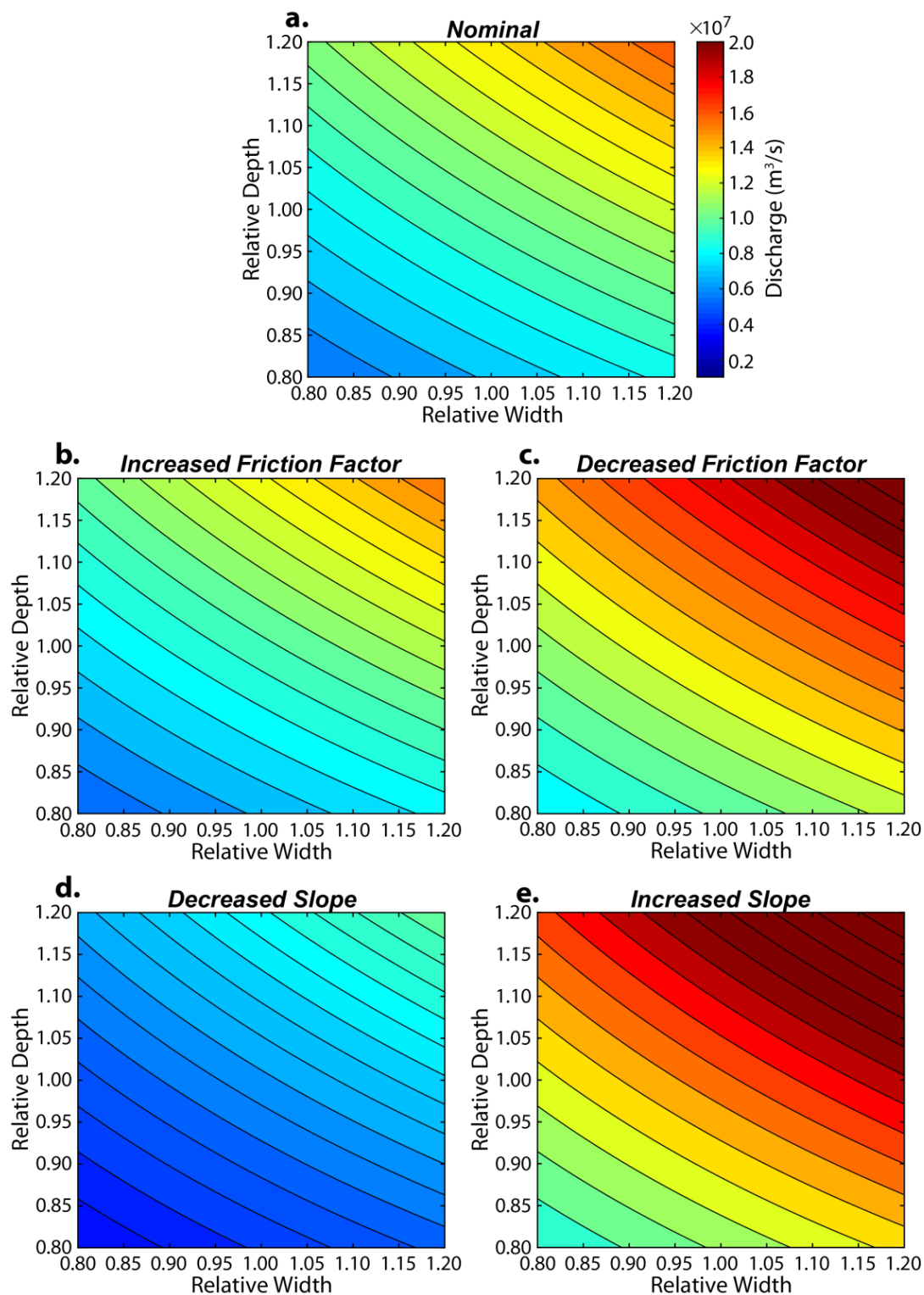


Figure 6. Upper Reull Vallis outflow channel system discharge estimate parameter sensitivity analysis. (a.) Discharge estimate for location 3 of the upper Reull Vallis system (Figures 3-5) as a function of channel depth and width relative to the measured

values accounting for a measurement error factor of ± 0.2 . Discharge is calculated using nominal values of channel slope and friction factor. **(b.)** Discharge estimates calculated with an increased friction factor assuming a boulder bed with a D_{84} of 0.25 m. **(c.)** Discharge estimates calculated with a decreased friction factor assuming a sand bed with a D_{50} of 0.001 m. **(d.)** Discharge estimates calculated with a reduced slope of 0.15° , the minimum slope measured in this analysis of the Reull Vallis system. **(e.)** Discharge estimates calculated with an increased slope of 1° , approximately double the largest slope measured in this analysis of the Reull Vallis system. Note that increasing the friction factor and decreasing the slope cause reduced discharges, while decreasing the friction factor and increasing the slope result in increased discharge values. Over the conservative range of measurement error and parameter values considered, the discharge estimates of the upper Reull Vallis system do not vary by more than an order of magnitude from the base case estimate ($10^7 \text{ m}^3/\text{s}$), with minimum values of $\sim 10^6 \text{ m}^3/\text{s}$ and maximum values of $\sim 2 \times 10^7 \text{ m}^3/\text{s}$.

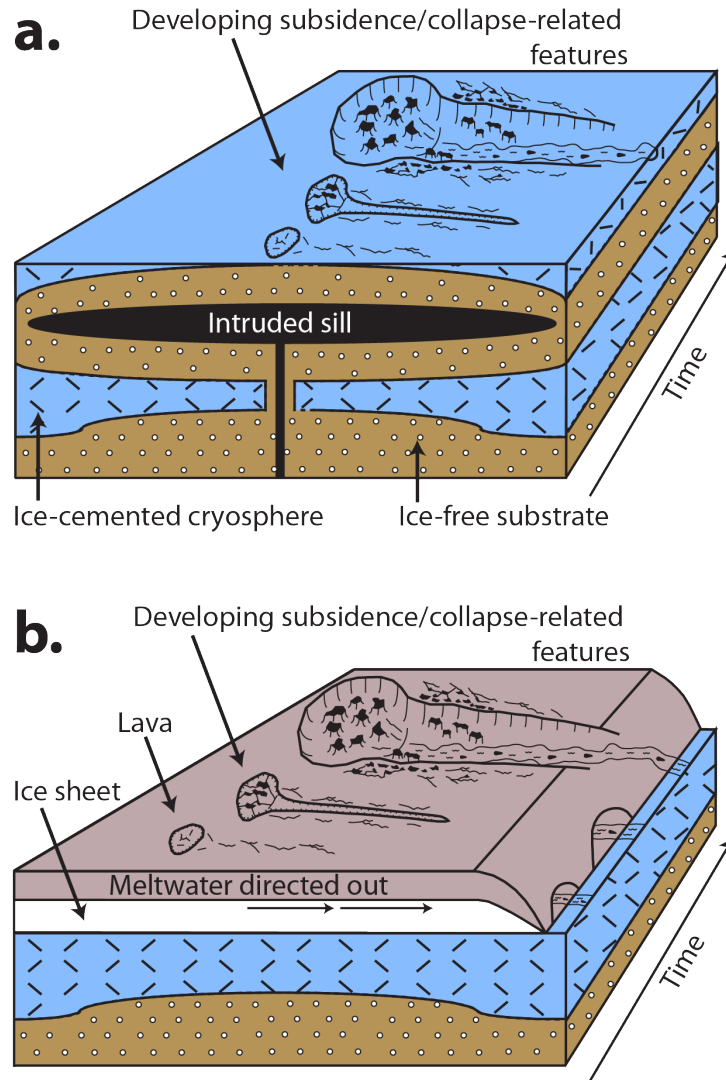
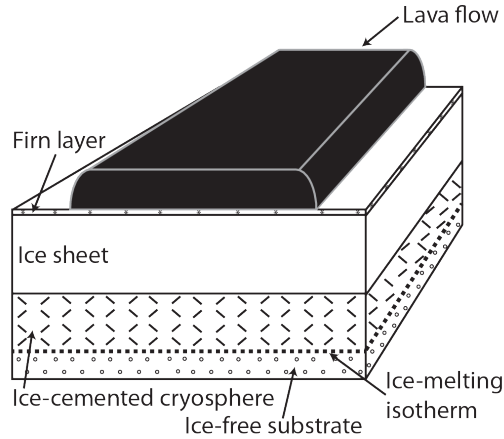


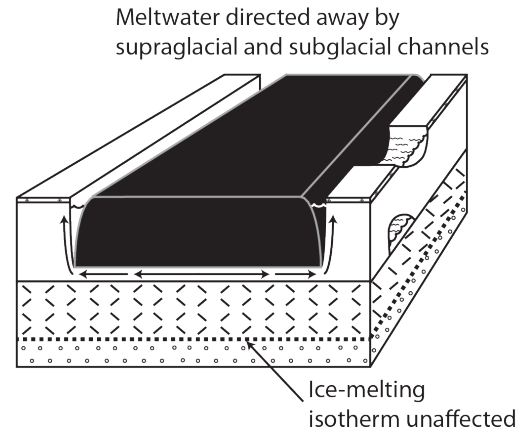
Figure 7. Models for the formation of the Hesperia Planum outflow channels. The systems contain source regions exhibiting varying degrees of collapse, a collapsed valley section, followed by the emergence of a fluvial valley in which flow features and erosional incision is evident. **(a.)** Outflow channel formation model invoking the melting of ground-ice through intrusive plutonic activity as previously proposed by Squyres et al., (1987). In this model heat transferred from the intruded sill causes melting of cryospheric ground-ice, which causes fluidization and mobilization of the regolith leading to collapse of the surface materials and the outflow of ice-cement meltwater. **(b.)** The proposed ice sheet lava-heating and loading model (modified from Cassanelli and Head, 2016) for outflow channel formation. Lava emplacement induces melting and removal of subsurface ice causing progressive collapse of the superposed lava. Meltwater is unable to infiltrate the impermeable ice (Cassanelli and Head, 2015; Cassanelli et al., 2015) and is directed along the ice-lava interface (Cassanelli and Head, 2016) causing fluvial erosion within the collapsed terrain.

Direct Top-Down Contact Melting

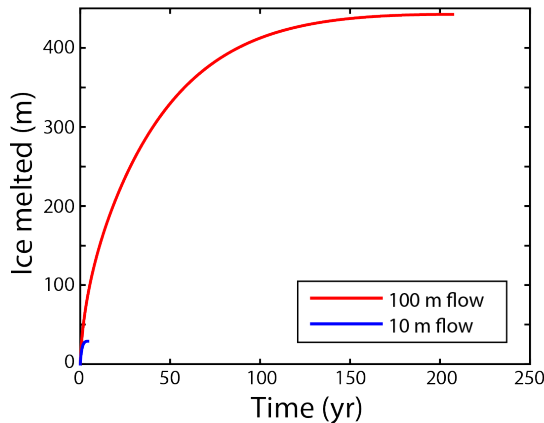
a. Initial lava emplacement



b. Contact heat transfer & top-down melting



c. Ice melted vs. time - variable flow thickness



d. Ice melted vs. time - lava flow sequence

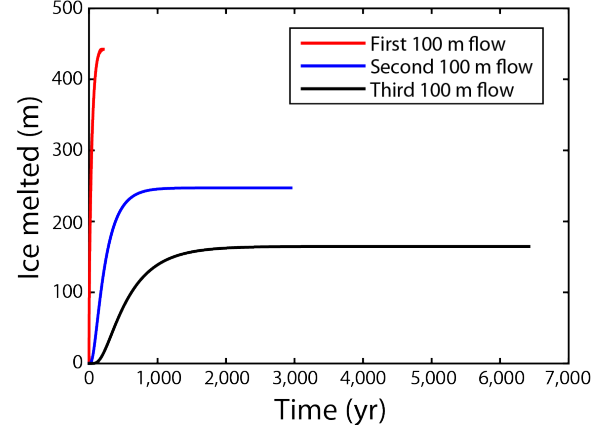
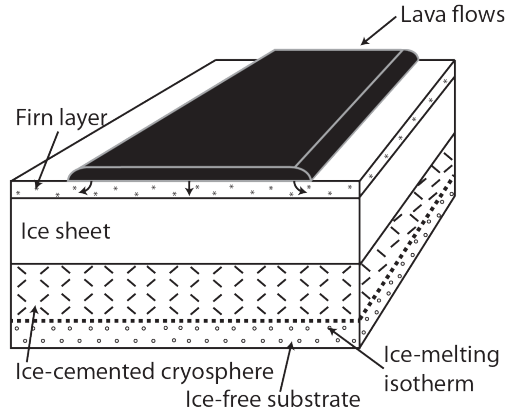


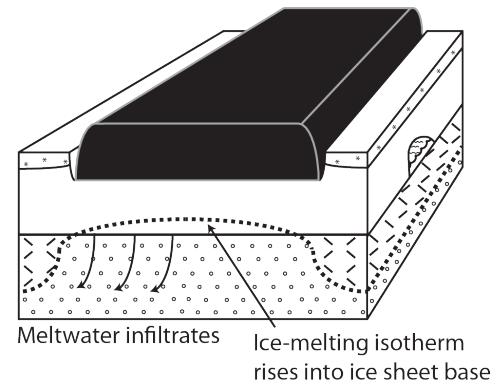
Figure 8. Summary diagram of the top-down melting processes induced by ice sheet lava heating and loading (figure and results adapted from Cassanelli and Head, 2016). The results displayed in panels (a.) and (c.) are from calculations performed by Cassanelli and Head (2016) which are improved upon in this work. (a.) Lava is initially emplaced upon an ice sheet exhibiting a thin firn layer with the ice-melting isotherm initially at depth within the subsurface producing an ice-cemented cryosphere (Cassanelli and Head, 2015). (b.) Heat is rapidly transferred through direct contact between lava and ice, causing ice sheet melting. The ice-melting isotherm at depth remains unaffected over the timescales associated with contact heating and melting. (c.) The relative thickness of ice melted as a function of time for lava flows of varying thickness. Thicker lava flows contain significantly more heat, and therefore produce much more melting over more prolonged periods of time. (d.) The relative thickness of ice melted as a function of time for a sequence of constructional lava flows. Successively emplaced lava flows produce less melting due to the thickening layer of previously emplaced lava inhibiting heat transfer downward into the ice.

Bottom-up Melting

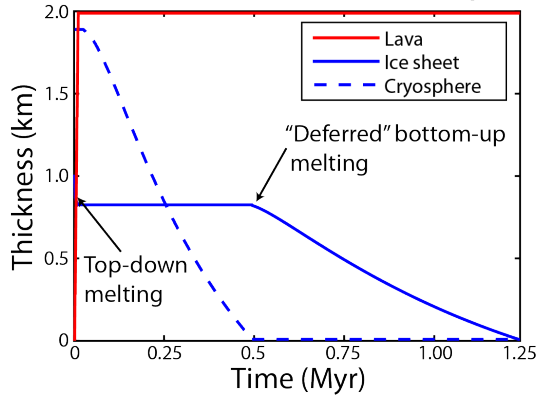
a. Initial lava emplacement



b. Bottom-up geothermal heating and melting



c. Geothermal-limited “deferred” bottom-up melting



d. Insulation-limited bottom-up melting

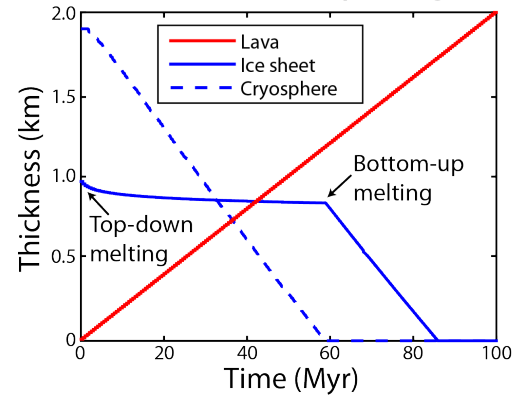


Figure 9. Summary diagram of the bottom-up melting processes induced by ice sheet lava heating and loading (figures and results adapted from Cassanelli and Head, 2016). (a.) Lava is initially emplaced upon an ice sheet exhibiting a thin firn layer with the ice-melting isotherm initially at depth within the subsurface producing an ice-cemented cryosphere. (b.) If the lava flows are thin, heat is inefficiently transferred into the ice, causing minimal top-down ice sheet melting. As the accumulated lava thickness increases, the ice sheet becomes insulated, raising the ice-melting isotherm from depth toward the ice sheet base inducing bottom-up melting of the cryosphere and eventually the ice sheet. Bottom-up melting rates are significantly less than top-down melting rates with melting occurring over much greater timescales. (c.) The thicknesses of the lava, ice sheet, and cryosphere layers during an ice sheet lava heating and loading scenario involving rapid lava accumulation and “deferred” bottom-up ice sheet melting. Lava is accumulated in 10 m increments over a rapid end-member total emplacement timescale of 10 kyr. (d.) The thicknesses of the lava, ice sheet, and cryosphere layers during an ice sheet lava heating and loading scenario involving prolonged lava accumulation leading to insulation-limited bottom-up ice sheet melting. Lava is again accumulated in 10 m increments, but over a maximum end-member total emplacement timescale of 100 Myr.

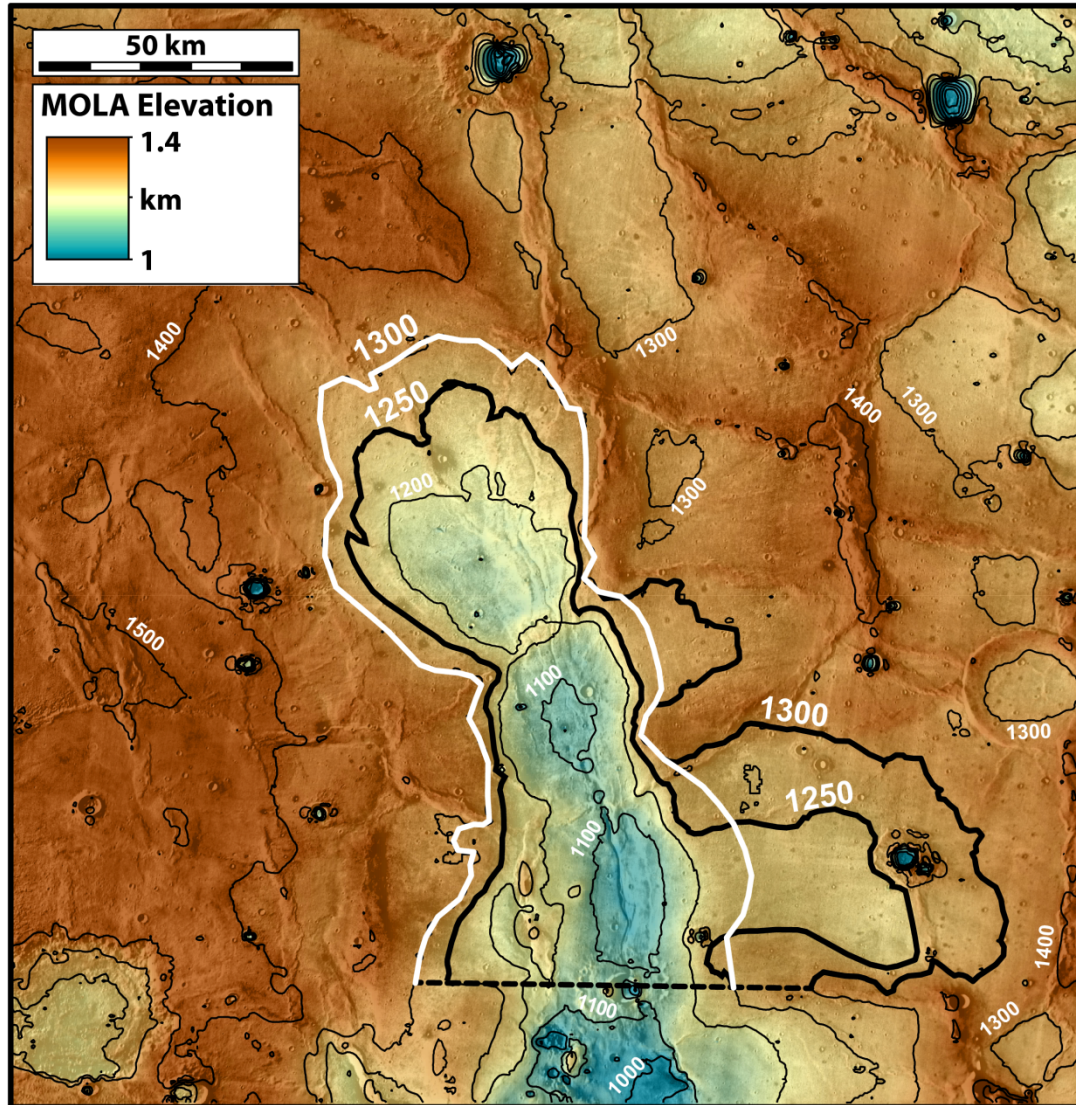


Figure 10. Contour map of the northern trough area of the Reull Vallis outflow channel system contoured in 100 m intervals. The 1250 m and 1300 m contours used to delineate the trough area are shown in bold, and the interpreted northern trough area is shown by the bold white line (covering an area of $\sim 6,100 \text{ km}^2$).

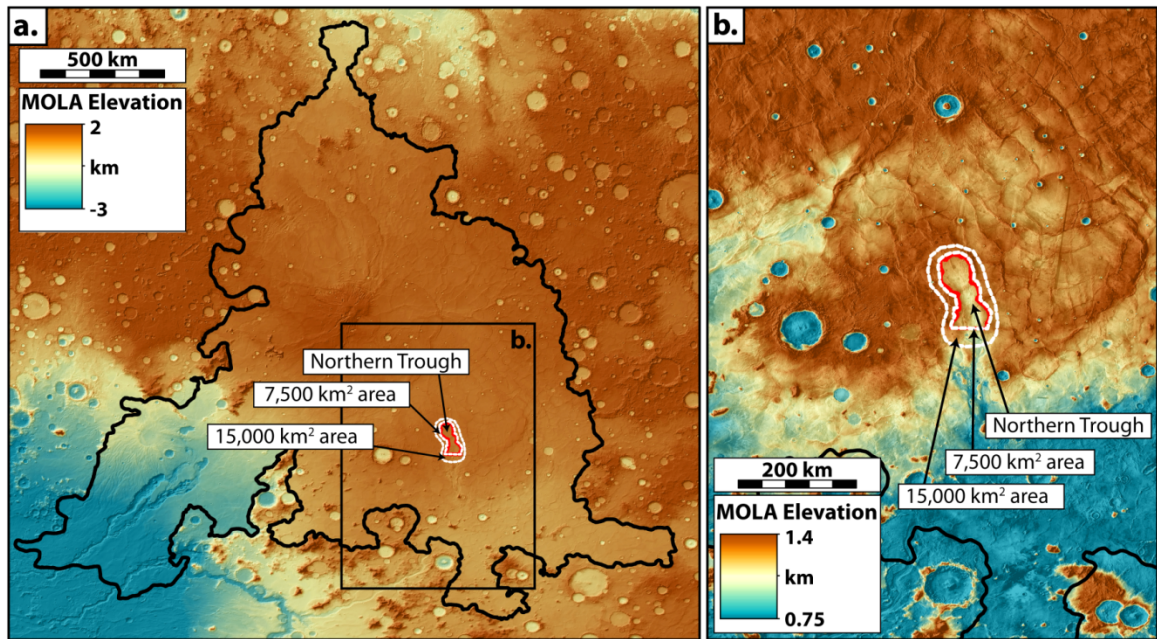


Figure 11. (a. and b.) Context images showing the extent of the areas over which melting from ice sheet lava heating and loading is required to provide the minimum volume of water necessary to form the upper Reull Vallis fluvial system.

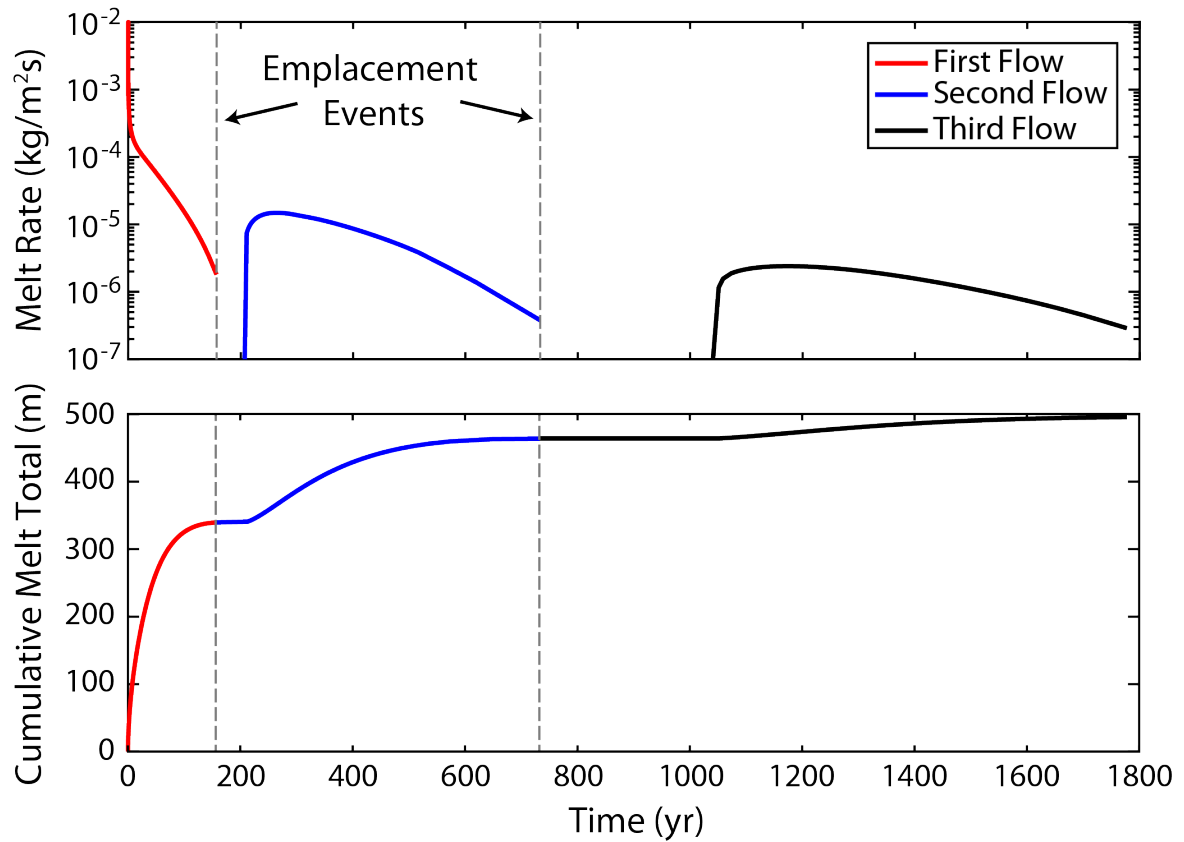


Figure 12. Melting rate estimates for an emplacement sequence of three 100 m thick lava flows atop an ice sheet. Successive lava flows are emplaced immediately following the cessation of melting from the preceding lava flow. Estimates are generated with the thermal model discussed in Section 4.2 and 4.3. Note that both melting rates and cumulative melt totals decrease rapidly with each sequential flow due to the presence of intervening cooled lava which reduces the efficiency of heat transfer from the heated lava to the ice.

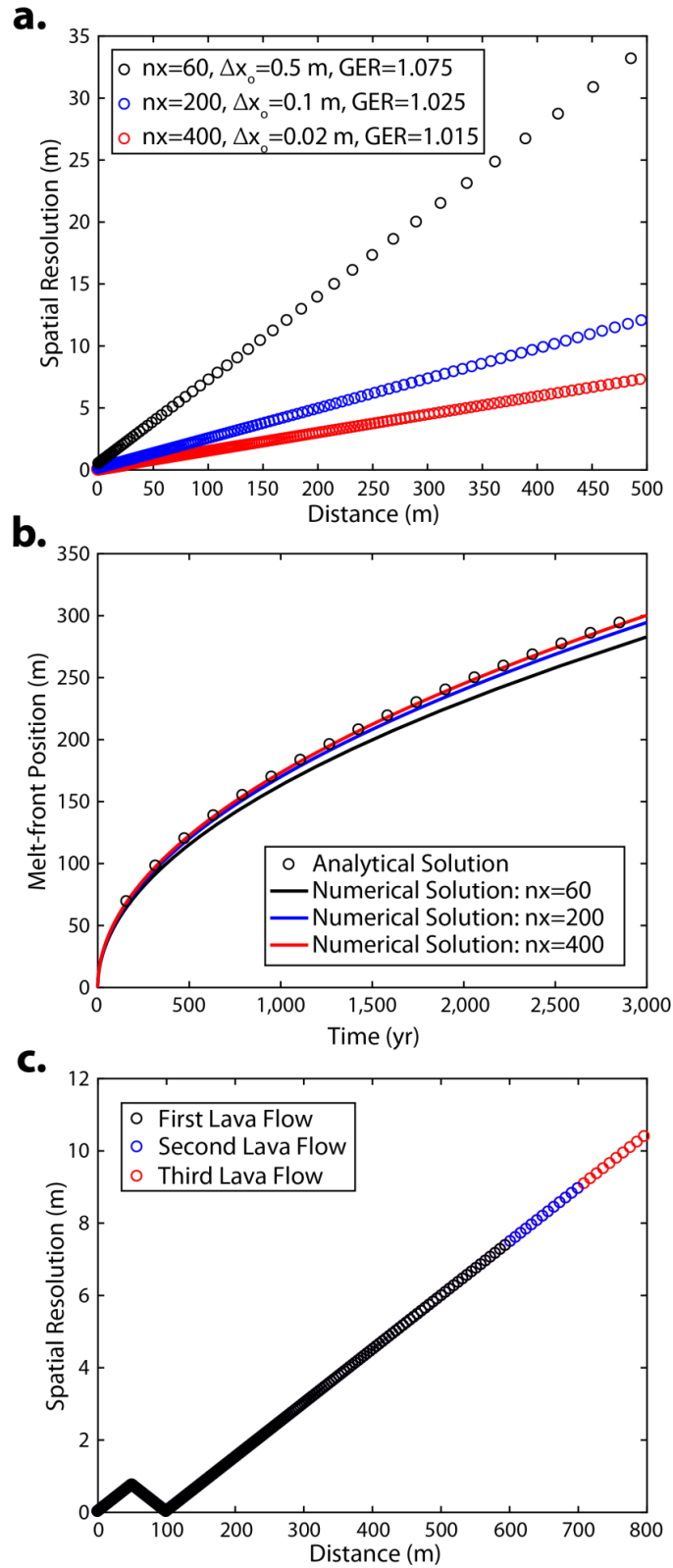


Figure 13. Phase-change numerical thermal model benchmarking analysis. The phase-change numerical thermal model developed and applied in this assessment was benchmarked against an analytical solution to a one-phase Stefan problem. **(a.)** Model

grid configurations used to perform the benchmarking analysis. The model grid construction process is described in Section 4.3. Numerical models were performed for a range of non-uniform grid configurations with progressively increasing spatial and temporal resolution using 60, 200, and 400 grid cells (nx), initial grid spacing values (Δx_o) of 0.5, 0.1, and 0.02 m, and grid expansion ratios (GER) of 1.075, 1.025, and 1.015. Time steps utilized in each model were set to a fraction of the minimum von Neumann stability criterion with values of 0.5, 0.1 and 0.04, respectively. **(b.)** Numerical model predictions of melt-front position versus time for the one-phase Stefan problem versus the analytical solution. At increased spatial and temporal resolution the numerical models converge to very close agreement with the analytical solution for the one-phase Stefan benchmark problem. **(c.)** Model grid configurations utilized for the ice sheet lava heating and loading melting analysis. Grid configurations between the model scenarios overlap but are extended to include additional layers of cooled lava in assessing melting induced by the second and third lava emplacement events.

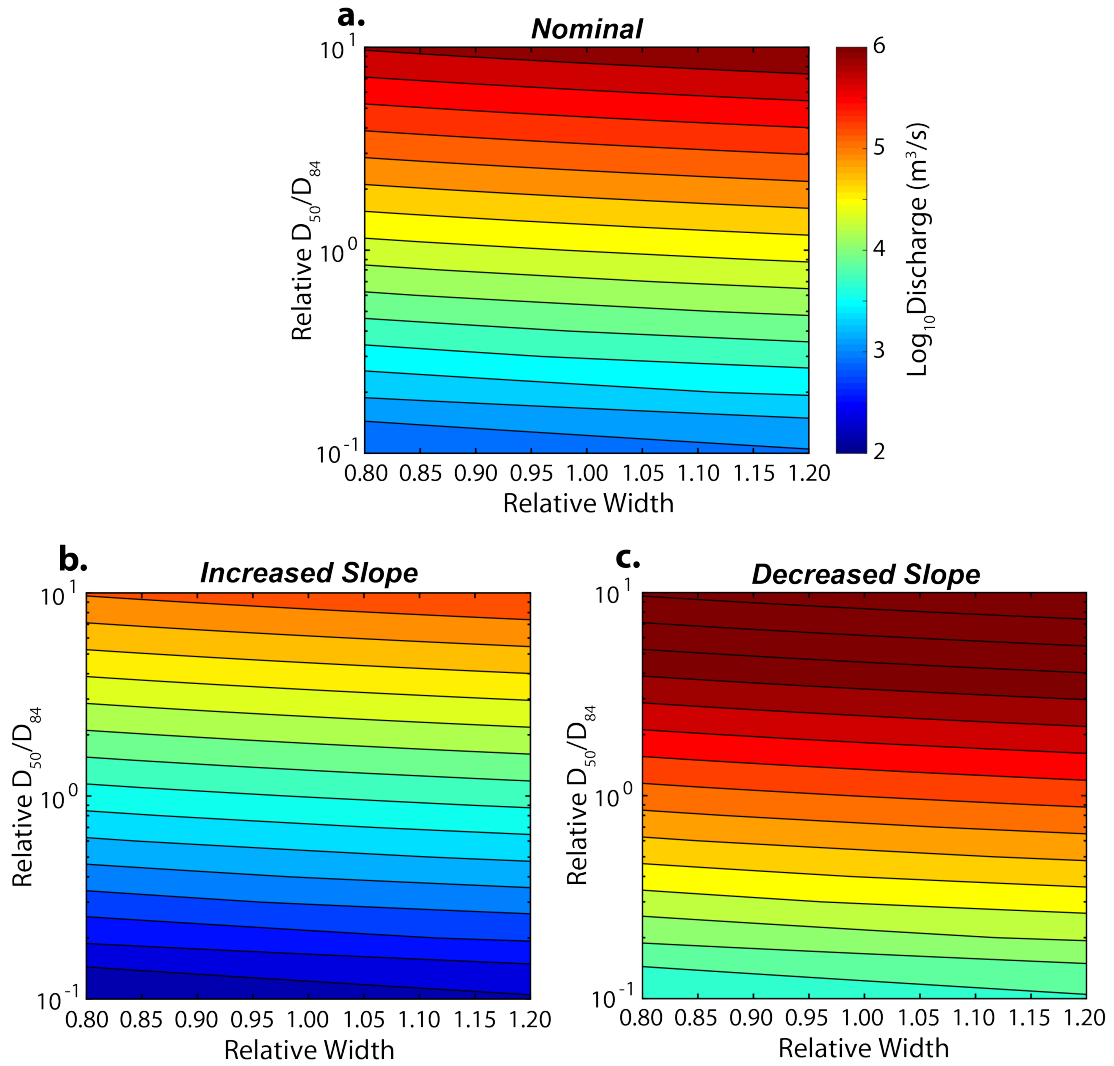


Figure 14. Parameter sensitivity analysis of the bed motion threshold minimum discharge estimate for the upper Reull Vallis system (detailed in Section 6). **(a.)** Discharge estimate for location 2 (Figures 3-5) of the upper Reull Vallis system as a function of the D_{50}/D_{84} (indicating relative values of both parameters, not their ratio) bed grain sizes and channel width relative to the nominal values accounting for an order of magnitude variation in the grain size parameters and a width measurement error factor of ± 0.2 . Discharge is calculated using the nominal measured value of channel slope (0.15°). **(b.)** Discharge estimates calculated with an increased channel slope of 1.5° . **(c.)** Discharge estimates calculated with a decreased channel slope of 0.015° . Note that the relative bed grain size exerts a stronger influence on the estimated discharges than the channel width, with increasing grain sizes requiring increased discharges to initiate bed motion. Increased channel slope decreases the discharges required to initiate bed motion for a given grain size.

**Chapter 3: Large-Scale Lava-Ice Interactions on Mars: Investigating its
Role During Late Amazonian Central Elysium Planitia Volcanism and
the Formation of Athabasca Valles**

James P. Cassanelli^a & James W. Head^a

^a Department of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

Published in: *Planetary and Space Science*

Cassanelli, J. P., and Head, J.W., (2018). Large-scale lava-ice interactions on Mars:
Investigating its role during late Amazonian Central Elysium Planitia volcanism and the
formation of Athabasca Valles. *Planetary and Space Science*, 158, 96-109, doi:
10.1016/j.pss.2018.04.024

Abstract

The Central Elysium Planitia region contains evidence for large-scale Late Amazonian volcanic and fluvial activity, the youngest on Mars. The region consists of a population of geologically recent, Late Amazonian-aged flood basalt volcanic plains which are incised by several large fluvial outflow channel systems. Recent understanding of the distribution of paleo-glaciation on Mars, informed by geomorphic indicators and model predictions, suggests that Central Elysium Planitia was an area of likely surface snow and ice accumulation throughout the Amazonian period. Furthermore, the volcanic plains of Central Elysium Planitia contain wide areas with preserved rootless cones, representing among the strongest evidence for past lava-ice interactions on Mars. Consequently, Central Elysium Planitia is of significant interest to understanding the recent Amazonian thermal and hydrologic history of Mars.

Here we perform an exploratory assessment of the influence of surficial large-scale lava-ice interactions on the geomorphology of the Central Elysium Planitia volcanic plains and their potential contribution to the formation of the fluvial outflow channels contained within the region. The outflow channels of Central Elysium Planitia are generally interpreted to have formed as a result of catastrophic outbursts of groundwater released from pressurized, confined subsurface aquifers by magmatic dike emplacement. However, the formation of the relatively young Central Elysium Planitia outflow channels by this mechanism is subject to several demanding conditions which are not required of a lava-ice interaction origin. Therefore we revisit the origin of the Central Elysium Planitia outflow channels and explore large-scale lava-ice interactions as a possible alternative mechanism to contribute to, or account for, their formation. This

assessment is performed through a case study of Athabasca Valles, a well-characterized outflow channel system which appears to emanate from the Cerberus Fossae fissure system.

We find that surface ice and lava-ice interactions could have influenced the geomorphology of Central Elysium Planitia by contributing to the formation of the regions platy-ridged volcanic plains texture and rootless cones. Large-scale lava-ice interactions in Central Elysium Planitia are found to be able to readily supply the water volumes needed to form the Athabasca Valles system, but additional mechanisms are required to cause meltwater confinement and release to supply requisite discharges. Of the several potential mechanisms which could have led to meltwater confinement and release, the most likely are englacial melt accumulation with outflow triggered by the flotation of a confining ice dam, and supra/englacial meltwater accumulation released by top-down supraglacial overflow. A simple analysis of these processes indicates that the rates of discharge necessary to form Athabasca Valles can be supplied while adhering to the geomorphic constraints of the system. Therefore, we conclude that the formation of the Athabasca Valles outflow channel system (and possibly other outflow channels in the region) could plausibly have been the result of surficial lava-ice interactions through meltwater confinement, accumulation, and release, providing an advantageous alternative to classical groundwater outburst models.

1. Introduction

The Central Elysium Planitia region of Mars (Figure 1) (*e.g.* Vaucher et al., 2009a, 2009b) (previously referred to as the Cerberus plains) contains evidence for the youngest volcanic and fluvial/diluvial activity on the martian surface (Plescia, 1990, 1993;

Vaucher et al., 2009a; Voigt and Hamilton, 2018) and is therefore of significant importance for understanding the thermal and hydrologic history of the planet. The region is characterized by a series of geologically recent, Late Amazonian-aged lava plains (Figure 1) which are interpreted to have been emplaced in a flood basalt mode (Plescia, 2003; Keszthelyi et al., 2004, 2017; Lanagan, 2004; Jaeger et al., 2010; Cataldo et al., 2014), involving voluminous eruptions of low viscosity magma at high effusion rates and with eruptive activity sustained over significant timescales (Head and Coffin, 1997). The flood basalt plains of Central Elysium Planitia are incised by the Athabasca, Grjota', Marte, and Rahway Valles outflow channel systems (Tanaka et al., 2014) (Figure 1). The Athabasca Valles system is of particular interest due to its exceptionally young apparent age, with crater age dating estimates suggesting fluvial activity may have occurred within the system as recently as the last ~20 Myr (Berman and Hartmann, 2002; Burr and McEwen, 2002). It is worth noting, however, that these crater retention age estimates are derived from the lava units which embay the Athabasca Valles channel, and thus, the channel itself could be much older (unless it was carved by the lava) and of unknown precise age (McEwen et al., 2012; Dundas and Keszthelyi, 2013). Despite these uncertainties, stratigraphic constraints (Voigt and Hamilton, 2018) generally indicate that the outflow channels of Central Elysium Planitia (Figure 1) are a relatively young, Middle to Late Amazonian-aged, population. It is widely accepted that the martian outflow channels were carved by water, thus the relatively young channel systems of Central Elysium Planitia (Figure 1) represent the most geologically recent large-scale fluvial activity on Mars.

Both the flood basalt plains of Central Elysium Planitia, as well as the water thought to have carved the outflow channels of the region, are inferred to have emanated from Cerberus Fossae (Figure 1 and 2) (Head et al., 2003b; Plescia, 2003; Jaeger et al., 2010), a series of extensional fissures extending over a total length of >1,000 km within the region. The Central Elysium Planitia outflow channels, as well as many other martian outflow channels, are canonically interpreted to have formed through the effusion of pressurized groundwater released from aquifers confined beneath the martian cryosphere (e.g. Baker et al., 1992; Clifford, 1993; Carr, 1996; Head et al., 2003b; Harrison and Grimm, 2004; Hanna and Phillips, 2005, 2006; Andrews-Hanna and Phillips, 2007; Marra et al., 2014). In Central Elysium Planitia, outflow channel formation by this mechanism is suggested to have been triggered by magmatic dike emplacement resulting in breaching of the cryosphere and outflow of pressurized groundwater (Head et al., 2003b) due to the close association of the outflow channel heads and the Cerberus Fossae fractures. This relationship is well-illustrated by the Athabasca Valles outflow channel (Figure 2) which unambiguously heads along the Cerberus Fossae. However, the importance of water in the formation of the Athabasca channel has been called into question (a concise summary is provided by Keszthelyi et al., 2017), with a range from purely fluvial to purely volcanic origins suggested, or through a combination of the two (Burr, 2003; Burr et al., 2005; Leverington, 2004; Keszthelyi et al., 2007, 2017; Jaeger et al., 2010; Ramsdale et al., 2015).

Assuming that the Central Elysium Planitia outflow channels were carved exclusively by water, the geologically young age of the systems (Voigt and Hamilton, 2018) results in several demanding requirements for the canonical formation mechanism. (1) A liquid

groundwater system must have been preserved at great depth beneath a very thick cryosphere to near the present day (against the relatively efficient action of vapor diffusive loss to the cryosphere; Clifford, 1991), and must have been supplied by a source of active groundwater recharge (most plausibly Elysium Mons, still an unlikely source of recharge late in the Amazonian period near the time of formation of the Elysium Planitia outflow channels; Russell and Head, 2007; Cassanelli et al., 2015). (2) The groundwater system must have been saturated to the base of the cryosphere to allow confinement and pressurization, requiring potentially significant volumes of liquid water (Clifford, 1993). (3) Dike emplacement or tectonic extension must have produced and maintained an open conduit across the entire ~5-10 km thick cryosphere (Clifford et al., 2010) to allow water outflow to the surface. (4) The permeability of the aquifer from which the floods emanated must have been unreasonably large, far exceeding those common of terrestrial aquifers at similar depths and scales (Head et al., 2003b).

Could any other mechanism account for the formation of the Central Elysium Planitia outflow channels without incurring these stringent requirements? Recent work has proposed that large-scale surficial lava-ice interactions could serve as a potential alternative mechanism for outflow channel formation (Cassanelli and Head, 2016, 2018), a mechanism which obviates many of the difficult conditions required of the canonical outflow channel formation model. Given the abundant evidence for lava (Vaucher et al., 2009a), ice (both surface and ground) (Head et al., 2003a; Mellon et al., 2004; Kadish et al., 2010; Bramson et al., 2015), and past lava-ice interactions (Hamilton et al., 2010b, 2011; Keszthelyi et al., 2010; Dundas and Keszthelyi, 2013) preserved in the Central Elysium Planitia region (or at comparable latitudes in the case of ice; Pedersen et al.,

2010; Pedersen and Head, 2011, 2010), this is a plausible mechanism to explain the formation of outflow channels in the region. Additionally, the coincidence of glaciation and volcanism in Central Elysium Planitia and the operation of large-scale lava-ice interactions could have resulted in significant modification of the volcanic plains and meltwater generation that may have contributed to shaping the geology of the region (Cassanelli and Head, 2016, 2018).

Here we perform a first-order evaluation of possible large-scale surface lava-ice interactions in Central Elysium Planitia. We explore the potential involvement of lava-ice interactions in producing the geomorphology of the Central Elysium Planitia volcanic plains and in the formation of the outflow channels of the region. We first review the nature and characteristics of volcanism and glaciation in Central Elysium Planitia as well as the preserved evidence for past lava-ice interactions in order to outline conditions of local lava and ice emplacement. We then review the geomorphology of Central Elysium Planitia volcanic plains and assess the possible influences of large-scale lava-ice interactions. Finally, we explore the formation of fluvial features in Central Elysium Planitia through large-scale lava-ice interactions. This final assessment is performed through a case study of Athabasca Valles (Figure 1 and 2) as it is the most intensely studied and best characterized outflow channel system of the region, and because it poses the most rigorous test of the possible lava-ice interaction origin due to its scale, channel head morphology, and proximity to the Cerberus Fossae source (later discussed in detail). The geomorphology of Athabasca Valles is first reviewed in order to outline required criteria that must be satisfied to account for formation of the channel. These criteria are then used to test against predictions of meltwater generation and geomorphology from

previous large-scale lava-ice interaction modeling work (Cassanelli and Head, 2016, 2018).

2. Lava-Ice Interaction Geometry

In our analysis of large-scale surficial lava-ice interactions in Central Elysium Planitia, we assume a lava-ice interaction geometry involving the emplacement of effusive lava flows atop regional snow and ice deposits, as proposed in the ice sheet lava heating and loading mechanism (Cassanelli and Head, 2016, 2018). While other lava-ice interaction geometries are possible, the supra-ice emplacement geometry provides testable maximum end-member estimates for meltwater generation by optimizing heat transfer to the surface ice (Head and Wilson, 2002, 2007; Wilson and Head, 2002, 2007; Cassanelli and Head, 2016, 2018) and has been observed to be a dominant interaction geometry between lava flows and surface snow and ice on Earth (Edwards et al., 2012, 2013, 2014, 2015). Supra-ice emplacement of lava flows can occur by (1) direct advancement of lava flows onto the ice from topographically-elevated terrain (as has been observed in glaciated terrestrial volcanic settings; Edwards et al., 2012, 2013, 2014, 2015), (2) supraglacial eruptions fed by dikes emplaced through surface ice due to high strain rates (Head and Wilson, 2002, 2007; Wilson and Head, 2002, 2007) or, (3) eruptions from tuyas that have pierced through surface ice (Smellie, 2007; Kadish et al., 2008; Scanlon et al., 2014).

3. Central Elysium Planitia: Regional Volcanism

The volcanic plains of the Central Elysium Planitia region are interpreted to have been emplaced in a flood basalt volcanic mode (Plescia, 2003; Keszthelyi et al., 2004, 2017; Lanagan, 2004; Jaeger et al., 2010; Cataldo et al., 2014). This style of volcanism is

characterized (Keszthelyi et al., 2000, 2004) by very voluminous eruptions of low viscosity magma at high effusion rates producing large, extensive sheet lava flows with eruptive activity sustained over significant timescales, similar to that of the large igneous provinces on Earth (Head and Coffin, 1997). Volcanism of this nature is responsible for the emplacement of volcanic plains across nearly 1/3 of the surface of Mars (Head et al., 2002) with minimal evidence for associated source vents (Gregg, 2017) and only very localized volcanic constructs (primarily in Hesperia Planum and Syrtis Major; Tanaka et al., 2014). Thus, the predominant mode of volcanic emplacement in Central Elysium Planitia is extrusive in character, without evidence for a major accompanying component of intrusive or explosive activity.

Within the Central Elysium Planitia region (Figure 1) (Plescia, 1990, 2003; Keszthelyi et al., 2000, 2004; Sakimoto and Gregg, 2001; Berman and Hartmann, 2002; Burr et al., 2002b, 2002a; Fuller and Head, 2002; Werner et al., 2003; Lanagan, 2004; Vaucher et al., 2009a, 2009b; Jaeger et al., 2010) the volcanic plains cover a total area of $\sim 1-10 \times 10^6 \text{ km}^2$ with a volume of $\sim 0.1-0.15 \times 10^6 \text{ km}^3$ and with typical individual lava flow unit thicknesses of $\sim 1-40 \text{ m}$, areas of $\sim 10^4 \text{ km}^2$, and volumes of $\sim 10^3 \text{ km}^3$. Crater retention ages (Vaucher et al., 2009a) of the Central Elysium Planitia volcanic plains suggest that volcanic activity within the region occurred as recently as $\sim 2 \text{ Ma}$ and extended to at least 250 Ma . Estimates suggest individual periods of volcanic activity within Central Elysium Planitia lasted on the order of 1 Myr (Vaucher et al., 2009a) with large portions of the province (specifically the $\sim 2.5 \times 10^5 \text{ km}^2$ Athabasca Valles flood lava) potentially emplaced over time periods as little as weeks to months (Jaeger et al., 2010).

Given these general characteristics, we evaluate large-scale lava-ice interactions in Central Elysium Planitia by assuming a total lava-ice interaction area of 10^7 km^2 (the maximum extent of the lava, also assuming equally extensive surface ice deposits) and individual lava flow thicknesses of 20 m and areas of 10^4 km^2 . In this analysis we assume that sequential lava flows are emplaced after the complete cooling of the prior flow, a reasonable assumption given the short duration of individual lava flow cooling relative to the total emplacement timescale of flood basalt units (Cassanelli and Head, 2016). Furthermore, we neglect the effects of the lava flow emplacement recurrence rate because, under these conditions, it does not control the total amount of ice melted, but only the long-term melting behavior by determining the amount of time in-between lava emplacement events and thus melt pulses (Cassanelli and Head, 2016, 2018).

4. Central Elysium Planitia: Snow and Ice

Several lines of evidence have been documented indicating that regional snow and ice were likely to have been common within the latitude range of Central Elysium Planitia during the Amazonian period of Mars history. These include: (1) The deposition and formation of tropical mountain glaciers on the Tharsis Montes and the ice-bearing latitude-dependent mantle resulting from redistribution of polar volatiles toward the equator during periods of increased obliquity (Kreslavsky and Head, 2002; Head and Marchant, 2003; Head et al., 2003a, 2005; Forget et al., 2006; Madeleine et al., 2009, 2014; Dundas et al., 2018). (2) The preservation and distribution of pedestal craters, interpreted to be direct indicators of paleo-ice sheet thickness, which exhibit heights of several tens to in excess of one hundred meters (Kadish et al., 2010; Kadish and Head, 2011; Fastook and Head, 2014). (3) The presence of ice-bearing lineated valley fill,

lobate debris aprons, and concentric crater fill units (Madeleine et al., 2009; Head et al., 2010; Levy et al., 2010; Fastook and Head, 2014). (4) General circulation model simulations of the Amazonian Mars climate predicting the accumulation of ice within similar latitudinal bands and within the northern portions of Central Elysium Planitia (Forget et al., 2006; Madeleine et al., 2009, 2014). (5) The inferred presence of widespread excess ground-ice within the nearby Arcadia Planitia region as indicated by Mars Reconnaissance Orbiter Shallow Radar (SHARAD; Seu et al., 2004) instrument measurements (Bramson et al., 2015).

Together, these observations and predictions suggest that the presence of regional surface snow and ice deposits in excess of tens of meters thick was likely within Central Elysium Planitia during the Amazonian period, coincident with the activity of volcanism in the region. Therefore, in our evaluation of lava-ice interactions in Central Elysium Planitia we assess a range of possible surface ice thicknesses from 1-100 m. This range accounts for a wide variation in possible surface ice thicknesses with the lower limit corresponding to thicknesses typical of the latitude dependent mantle (Kreslavsky and Head, 2002; Schon et al., 2009) and the upper limit to maximum thicknesses indicated by preserved pedestal craters (Kadish et al., 2010).

5. Central Elysium Planitia: Lava-ice Interactions

The volcanic plains of Central Elysium Planitia are not only of great interest due to their recent formation, but also because they contain among the best evidence for past lava-ice interactions on Mars. This evidence is in the form of rootless cones (also referred to as pseudocraters) which are preserved in numerous locations across the region, and which have been the subject of considerable study (Greeley and Fagents, 2001; Lanagan

et al., 2001; Burr et al., 2002a; Fagents et al., 2002; Bruno et al., 2004, 2006; Fagents and Thordarson, 2007; Hamilton et al., 2010b, 2011; Keszthelyi et al., 2010, 2017; Dundas and Keszthelyi, 2013; Cataldo et al., 2014; Marcucci et al., 2016). On Earth, rootless cones are known to form when lava flows advance over volatile-bearing materials (due to the generation and accumulation of steam to explosive pressures beneath the lava flow; Dundas and Keszthelyi, 2013), typically water-laden sediments and ice-bearing permafrost (Thorarinsson, 1953; Fagents and Thordarson, 2007; Hamilton et al., 2010a, 2010c; Belousov et al., 2011; Marcucci et al., 2016; Smellie and Edwards, 2016). A possible interpretation of the Central Elysium Planitia rootless cones, given the typical terrestrial formation conditions, is that the volcanic plains were emplaced onto water-laden regolith produced as a result of recent prior outflow channel activity. Consequently, this interpretation suggests a geologically contemporaneous age of fluvial and volcanic activity, therefore indicating a young geologic age for the large-scale outflow channel fluvial activity. However, modeling of rootless cone formation on Mars (Dundas and Keszthelyi, 2013) indicates that cones may have formed by the vaporization of shallow, atmospherically-emplaced ground-ice, which is known to be present across Mars (Mellon and Jakosky, 1995; Mellon et al., 1997, 2009), and which could have long predated volcanic activity (though the ability of rootless cones to form in this way depends on the depth of the permafrost ice-table which, if large, may have been difficult). Therefore, the rootless cones of Central Elysium Planitia are not definitive indicators of recent fluvial activity in the region (Keszthelyi et al., 2017).

In a scenario of large-scale surficial lava-ice interactions as investigated here, rootless cone formation would be expected to occur by direct molten fuel-coolant interactions

(Dundas and Keszthelyi, 2013) during lava flow advancement over snow and ice, or by advancement of lava flows over water-laden regolith produced by meltwater (Figure 3a). While the direct advancement of lava flows over surface snow and ice has recently been documented on Earth (Edwards et al., 2012, 2014, 2015), phreatovolcanic activity was relatively limited and rootless cone formation was not observed. However, observations of this process are currently very limited, and specific local conditions (such as steep topographic gradients and permeable snow allowing for rapid drainage of meltwater preventing major steam creation) may have inhibited rootless cone formation in these particular events (Edwards et al., 2012). Experiments involving the flow of lava directly over ice (Edwards et al., 2013) are suggestive of the ability of lava flows to trap steam and meltwater when flowing directly over ice, and thus rootless cone formation during the advancement of lava over surface snow and ice (Figure 3a) remains a plausible process.

Despite uncertainties regarding the precise formation conditions, the extensive distribution of rootless cones within Central Elysium Planitia, (covering at least thousands of square kilometers; Hamilton et al., 2011) provides strong support for the widespread past occurrence of lava-ice interactions within the region and justification for the scenario we investigate. With the characteristics of regional volcanism and glaciation within Central Elysium Planitia outlined, and the history of lava-ice interactions established, we now evaluate the potential contribution of lava-ice interactions toward the geomorphology of the region and the formation of outflow channels.

6. Geomorphology of Central Elysium Planitia

The geomorphology of the Central Elysium Planitia surface is characterized predominantly by young volcanic plains (Plescia, 1990; Fuller and Head, 2002; Lanagan, 2004; Jaeger et al., 2007, 2010; Vaucher et al., 2009a, 2009b; Morgan et al., 2013, 2015; Tanaka et al., 2014; Voigt and Hamilton, 2018) exhibiting lobate scarps, sinuous troughs, ridges, and platy-terrains (Keszthelyi et al., 2004; Voigt and Hamilton, 2016, 2018). A geomorphic attribute of particular interest within Central Elysium Planitia is the platy-ridged terrain morphology (Figure 4) which is expressed across wide areas of the volcanic plains in the region (Keszthelyi et al., 2004) and which is now assessed in detail.

6.1 Platy-Ridged Terrain

The platy-ridged surface morphology (Figure 4) of Central Elysium Planitia volcanic plains is characterized (Keszthelyi et al., 2004) by: (1) ridges tens of meters in height and width and hundreds of meters in length, (2) ridge-bounded flat plates hundreds of meters to kilometers across, (3) smooth surfaces divided into several-meter wide polygons which are elevated toward the center, (4) 1-10 km long parallel grooves cut into the surface (interpreted to result from flow around obstacles; Keszthelyi et al., 2004), and (5) inflated margins (interpreted to be inflated pahoehoe margins; Keszthelyi et al., 2004). In some areas, the platy-ridged volcanic plains surface is divided by long, kilometer-scale (in both length and width) lineated grooves which separate the terrain into compound-plates comprised of multiple individual plates and ridges (Figure 4).

The platy-ridged surface morphology is not well represented by terrestrial analogs (Keszthelyi et al., 2000, 2004) and its formation has been the subject of some debate. Alternative suggestions for the formation of the platy-ridged morphology have included:

(1) the disruption of an initial pahoehoe surface by large surges in the lava effusion rate, resulting in rafting of a disrupted lava crust (Keszthelyi et al., 2000, 2004; Ramsdale et al., 2015), (2) a combination of mud and ice flowing over the surface (Williams and Malin, 2004), (3) remnants of an evaporated/sublimated frozen sea (Murray et al., 2005), and (4) settling of icebergs onto martian plains (Woodworth-Lynas and Guigné, 2003).

Given the likelihood of volcano-ice interactions in Central Elysium Planitia during the Amazonian period, as was earlier established, we assess here whether interactions between the effusive Central Elysium Planitia lava flows and surface snow and ice may have contributed to, or resulted in, the formation of the platy-ridged terrain.

6.2 Platy-Ridged Terrain: Lava-Ice Interaction Origin

Observations (Edwards et al., 2012, 2014, 2015) of the advancement of lava flows over snow and ice on Earth have demonstrated that lava flows can undergo post-emplacement subsidence due to melting of underlying snow and ice leading to fracturing and disruption of the lava flow surface. While the lava flow deformation observed in these terrestrial studies is not directly similar to that exhibited by the platy-ridged terrain on Mars (which may be due to differences in scale, as the predicted surface ice deposits and sheet lava flows of the Central Elysium Planitia volcanic plains are much greater in size than those documented in the current terrestrial studies), these observations demonstrate that it is, at a minimum, possible for surficial lava-ice interactions to cause lava flow disruption and deformation. Therefore, the formation of the platy-ridged terrain by a surficial lava-ice interaction mechanism is, to a first order, plausible.

In order for surficial lava-ice interactions to account for the formation of the platy-ridged terrain in detail, the mechanism must be able to explain the complete range of

morphologic characteristics (Figure 4) including: (1) the lack of associated phreatovolcanic features (rootless cones), (2) the low variability in surface elevation between plates (on the order of 1-10 m), (3) the occurrence of raised ridges between plates, and (4) the separation of plates by lineated grooves up to ~1 km or greater in width.

Phreatovolcanic features: It is reasonable to expect the formation of platy-ridged terrain by surficial lava-ice interactions to be accompanied by considerable phreatovolcanic activity leading to rootless cone formation due to molten-fuel coolant interactions between lava and ice (Figure 3a). However, rootless cones are not consistently observed in conjunction with the platy-ridged terrains in Central Elysium Planitia. The lack of associated phreatovolcanic features can be accounted for by steam escape through the disrupted lava flow surface and margins as well as meltwater escape from beneath the lava flow through either absorption into underlying snow/ice or drainage beyond the flow margins. Both of these processes inhibit the accumulation of explosive steam pressures beneath the lava and have been observed (Edwards et al., 2012, 2014, 2015) to hinder phreatovolcanic activity and rootless cone formation during the advancement of lava flows over surface snow/ice on Earth. Therefore, the lack of phreatovolcanic features in association with the platy-ridged terrain does not preclude a lava-ice interaction origin.

Elevation variability: Differences in elevation between plates of the platy-ridged volcanic terrain are generally uniform and on the order of 1-10 m in magnitude. In a lava-ice interaction origin for the platy-ridged terrain, differential settling due to surface ice melting is required to cause lava flow surface disruption. Therefore, the observed 1-10 m

elevation differences between plates suggest that surface snow and ice must have been at least equal to these thicknesses in order for the platy-ridged texture to have formed by lava-ice interactions. Additionally, the existence of coherent lava plates several kilometers in dimension (Figure 4) indicates the plates did not subside over significant distances, as differences in settling rate over a large total amount of subsidence would likely cause fracturing and disruption of the surface to a finer scale. Therefore, if indeed the platy-ridge terrain is the result of surficial lava-ice interactions in Central Elysium Planitia, surface ice thicknesses must have been on the order of ~1-10 m, which is in general agreement with estimates of surface ice thickness derived from the latitude-dependent mantle (Kreslavsky and Head, 2002) and pedestal crater heights (Kadish et al., 2010).

Inter-plate ridges and lineated grooves: The ridges that bound the plates of the platy-ridged terrain (Figure 4) are interpreted to be pressure-ridges formed through the compression of adjacent lava plates (Keszthelyi et al., 2004). These pressure ridges are contrasted in some areas by lineated grooves which separate the lava plates and compound-plates of the platy-ridged terrain by widths up to ~1 km (Figure 4). The lineated grooves are thought to be produced as counterparts to the pressure ridges, with formation occurring in areas where the chilled lava surface underwent extensional flow rafting the lava plates apart and opening the grooves between. Therefore, the interpreted origin of the ridges and grooves suggests that their formation must have occurred during lava flow emplacement while the lava was still fluid and undergoing active flow.

In a lava-ice origin scenario, pressure ridges and grooves may have been formed as a result of differential ice sheet melting rates or variations in ice sheet surface elevation

leading to lava flow convergence toward areas of excess melting or lowered surface elevation and divergence in areas of reduced melting or increased surface elevation (Figure 3b). In order for the formation of ridges and grooves to occur in this manner the timescale for viscous flow of the lava must not exceed that of the diffusive cooling to enable the lava to adjust to induced potential changes through flow before solidification. This can be confirmed in a very simple manner by taking a ratio of the timescale required for thermal diffusion across the lava flow thickness (assuming lava flow cooling predominantly by conduction and neglecting advective effects) and viscous flow over a given distance (assuming Newtonian rheology for simplicity) which yields:

$$\frac{2 L \kappa \mu}{H^4 \rho g \sin \alpha} \quad (1)$$

where L is a flow distance of interest (~ 1 km in this case, an approximate average width of the lineated grooves and the greatest flow distance required to explain platy-ridge formation given that ridges are only tens of meters in elevation and hundreds in width), κ is the thermal diffusivity (for which a typical value of 10^{-6} m²/s is adopted), μ is the lava viscosity (for which a value of 10^4 Pa s is used, consistent with measurements from terrestrial basaltic lava flows; Harris and Allen, 2008), H is the lava flow thickness (for which the average 20 m value is utilized), ρ is the lava density (assuming a typical value of 3,000 kg/m³), g is gravitational acceleration (3.71 m/s² on Mars), and α is the slope (for which a value of 1° is used, approximately equivalent to a surface elevation change equal to the 20 m thickness of the lava flow over the typical 1 km length of the groove separations). Substituting the relevant parameters indicates a ratio value on the order of 10^{-7} , suggesting that, for the conditions of interest, viscous flow occurs over more rapid

timescales than lava flow cooling, supporting the ability of surficial lava-ice interactions to form the ridges and grooves (Figure 3b).

It is worth noting that, while surficial lava-ice interactions may have been responsible for the formation of the ridges and grooves of the platy-ridged lava terrain, they are not necessarily required. Formation of the ridges and grooves could alternatively have resulted from variations in lava flow conditions (as previously proposed; Keszthelyi et al., 2000, 2004) during supra-ice lava flow emplacement without any direct influence from differential settling due to melting of underlying surface snow and ice. This is because the conductive melting of ice generally proceeds at a slower rate than lava flow emplacement (Edwards et al., 2012, 2013, 2014, 2015), particularly for effusive low viscosity flows, such as those emplaced in Central Elysium Planitia (Keszthelyi et al., 2000, 2004; Jaeger et al., 2010). Ridges and grooves formed during emplacement in this manner could then have been preserved through subsequent melting and subsidence in areas where surface snow and ice was thin, and thus the total amount of subsidence experienced by the flow was limited. Therefore, platy-ridged terrain formation in this manner can occur irrespective of contemporaneous surface snow and ice deposits instead of as a direct consequence of lava-ice interactions.

Distribution: A final morphologic factor which requires consideration is the observed (Keszthelyi et al., 2000, 2004) occurrence of platy-ridged terrains on martian flood lavas outside of Central Elysium Planitia. The formation of platy-ridged flood lavas by lava-ice interactions as described in the preceding assessment (Figure 3b) necessitates the presence of surface snow and ice. Therefore, if indeed platy-ridged terrain formed in this manner (Figure 3b), the presence of this morphologic texture on the martian flood lava

plains could provide information on the ages (through crater age dating of lava flow surfaces), thicknesses, and distribution of past regional surface snow and ice deposits on Mars.

To summarize, platy-ridged terrain (Figure 4) may have formed as either a direct result of surficial lava-ice interactions (Figure 3b) or from variations in lava flow conditions during supra-ice emplacement without substantial influence from melting of underlying ice. In either case, however, the presence of surface snow and ice may have resulted in enhanced lava flow velocities and emplacement rates through the formation of a lubricating meltwater lens produced through melting of underlying snow and ice as documented in terrestrial experiments (Edwards et al., 2013). This process may help explain the remarkable length of the Central Elysium Planitia lava flows which, despite relatively low thicknesses (~20 m for individual flows) and exceedingly low regional slopes (~0.02-0.04%; Kreslavsky and Head, 2000), extend over lengths of at least 1,500 km and likely greater than 2,000 km (Keszthelyi et al., 2000, 2004). The significant lengths of the Central Elysium Planitia flood basalts have been previously explained by low lava viscosities, high effusion rates, and by lava flow emplacement occurring largely through flow confined beneath a protective cooled upper crust (increasing the thermal efficiency of the lava flow) (Keszthelyi et al., 2000, 2004). Thus, as earlier noted, if surface snow and ice was directly involved in the formation of platy-ridged terrain (through lava-ice interactions; Figure 3b), or indirectly by enabling greater lava flow emplacement rates, the platy-ridged terrain could provide information on the age, thicknesses, and distribution of past glacial activity on Mars. In addition to the influences lava-ice interactions may have had on the geomorphology of the Central Elysium Planitia

lava flows, the potential remains for significant meltwater production that may have contributed to the formation of fluvial features in the region. In order to test this we now examine the possible role of surface lava-ice interactions in the formation of the Athabasca Valles outflow channel system within the region.

7. Outflow Channel Formation by Lava-ice Interactions

Formation of outflow channels by large-scale interactions between lava and surface snow/ice is predicted to proceed by meltwater generation and release from both direct top-down contact heating and melting, and by insulation-induced bottom-up heating and melting (Cassanelli and Head, 2016, 2018). Here we consider only direct top-down contact heating and melting as insulation-induced bottom-up heating and melting would require the accumulation of significant thicknesses of lava (~2 km, far exceeding the estimated total thicknesses of the volcanic plains in Central Elysium Planitia) to initiate bottom-up melting (Cassanelli et al., 2015; Cassanelli and Head, 2015, 2016) for even the thickest ice sheets likely to have accumulated in the Central Elysium Planitia region during the Amazonian (~100 m, as indicated by the heights of pedestal craters at similar latitudes; Kadish et al., 2010). Lava-ice interactions involving top-down contact heating and melting (as envisioned in the ice sheet lava heating and loading model; Cassanelli and Head, 2016, 2018) depend upon a number of factors including: (1) the thickness of the individual accumulating lava flows, (2) the total thickness of lava accumulated, (3) the rate of lava accumulation, (4) the area over which top-down heating and melting occurs, and (5) the thickness and extent of pre-existing ice. These critical parameters and conditions for the Athabasca Valles outflow channel system were earlier defined through

reviews of the geomorphology of the Central Elysium Planitia region (Sections 2-4) and are now utilized to explore the formation of the outflow channel system.

7.1 Geomorphology of Athabasca Valles

To evaluate the ability of lava-ice interactions in Central Elysium Planitia to contribute to outflow channel formation in the region, we examine the Athabasca Valles system (Figure 1 and 2) as a case example. Of the outflow channel systems within Central Elysium Planitia, Athabasca Valles presents the most difficult test of the lava-ice interaction origin mechanism. This is because the Athabasca Valles system is (1) among the largest in the region, requiring substantial volumes and discharges of liquid water, and, (2) is the most proximal to the Cerberus Fossae lava source, thereby providing the least amount of possible lava-ice interaction area for meltwater generation and drainage collection. The Athabasca Valles outflow channel system is situated in the north of Central Elysium Planitia, with a 10-30 km wide main branch that extends over a distance of ~300 km from Cerberus Fossae toward the southwest (Figure 1 and 2). The main branch of the outflow channel system exhibits an average downstream slope of ~0.05%, varies in depth from ~60-75 m (shallowing beyond detection at the distal end), and is more deeply incised within the headwaters near the Cerberus Fossae source area (Burr et al., 2002a; Keszthelyi et al., 2007). Peak discharge estimates of $1\text{-}2 \times 10^6 \text{ m}^3/\text{s}$ have been derived from measurements of the main branch of the Athabasca Valles system (Burr et al., 2002b) and have been shown to be consistent with the morphology of the channel through subsequent flood modeling (Keszthelyi et al., 2007; Kim et al., 2014). Conservative estimates of the total volume of water required to erode the Athabasca Valles outflow channel system range from 400-1,600 km³ (Hanna and Phillips, 2006).

These discharge and volume estimates imply erosion of the Athabasca Valles system over a timescale of ~2-18 days assuming continuous, uninterrupted flow at the peak discharge values, comparable to those derived for other martian outflow channel systems under the same assumptions (Carr, 1996; Leask et al., 2006, 2007; Turbet et al., 2017).

These outlined geomorphic characteristics lead to important criteria that must be met in order to account for the formation of the Athabasca Valles outflow channel. (1) Supply of a total volume of 400-1,600 km³ of water as indicated by the total volume of the channel and conservative assumptions of the required water-to-rock ratio for erosion (Hanna and Phillips, 2006). (2) Supply of liquid water at discharge rates of ~1-2x10⁶ m³/s as evidenced by the channel dimensions (Burr et al., 2002b) and flood modeling (Keszthelyi et al., 2007; Kim et al., 2014). (3) Greater incision of the channel near the Cerberus Fossae source area as indicated by the channel morphometry (Burr et al., 2002a; Keszthelyi et al., 2007).

7.2 Origin of Athabasca Valles by Lava-Ice Interactions

The formation of Athabasca Valles by lava-ice interactions, and the conditions required to meet the geomorphic criteria of the system, are assessed through the implementation of models and predictions developed in previous investigations of lava-ice interactions on Mars (Cassanelli and Head, 2016, 2018). In particular, we utilize thermal models (Cassanelli and Head, 2016, 2018) to evaluate the process of top-down contact melting following supra-ice lava flow emplacement to estimate the magnitudes and rates of potential melting. Estimates of the volumes and rates of meltwater production are then used to evaluate the areas of melting required to produce the needed volumes and discharges to form the Athabasca Valles outflow channel.

The accumulation of 20 m thick individual lava flows (the typical value as established in Section 3) to the total thickness of tens to hundreds of meters observed (Vaucher et al., 2009a) in the Central Elysium Planitia region can potentially melt hundreds of meters of ice (Cassanelli and Head, 2016). Given that a plausible maximum surface ice sheet thickness in the Central Elysium Planitia region is ~ 100 m (as indicated by pedestal crater heights at similar latitudes; Kadish et al., 2010), the total melting volume able to be supplied by lava-ice interactions is likely to be limited by the amount of surface ice present. Therefore, we assume any surface ice present within Central Elysium Planitia to have undergone complete melting in response to the emplacement of the volcanic plains. The melting area required to supply the volume of water needed to form Athabasca Valles is then computed using a plausible range of surface ice thicknesses (as determined in Section 4) from 1-100 m (Figure 5). At a minimal surface ice thickness of 1 m, a total melting area of $\sim 0.4\text{-}1.7 \times 10^6 \text{ km}^2$ is required (Figure 5) to supply the range of water volumes estimated to have been needed to carve the Athabasca Valles system ($400\text{-}1,600 \text{ km}^3$). This area is approximately equivalent to the total area of the Central Elysium Planitia lavas and therefore, in principle, melting of 1 m of regional ice by lava-ice interactions could have supplied the required volumes of water to source the outflow channel. However, this would require the unlikely collection and focus of meltwater from across the entire region to the Athabasca Valles channel head. At the upper limit 100 m surface ice thickness the melting area requirement is reduced $\sim 0.4\text{-}1.7 \times 10^4 \text{ km}^2$ (Figure 5). This melting area is comparable to the average areal extent of individual lava flows in Central Elysium Planitia (refer to Section 3 for discussion on lava flow areas), indicating

that if surface ice were present at these thicknesses, lava-ice interactions could readily supply the water needed to form Athabasca Valles (Figure 5).

To assess the ability of lava-ice interactions in Central Elysium Planitia to supply meltwater at the discharge rates required to form Athabasca Valles we evaluate the melting rates (using previously developed lava-ice interaction thermal models; Cassanelli and Head, 2016, 2018) generated by the emplacement of a sequence of two 20 m thick lava flows (the typical thickness of individual lava flows in the region). Melting rates generated by the emplacement of additional lava flows are not considered as two 20 m thick lava flows are sufficient (Cassanelli and Head, 2016) to melt the maximum thickness of ice that is likely to have been present in Central Elysium Planitia (~ 100 m as discussed in Section 4). Results of this analysis predict peak top-down melting rates of $\sim 10^{-3}$ kg/m²s and average rates of $\sim 10^{-4}$ kg/m²s over the course of lava flow emplacement and heat transfer to the underlying ice (Figure 6). At these melting rates, top-down contact melting induced by lava-ice interactions would need to occur over areas of $\sim 10^6$ to $>10^7$ km² to supply the nominally estimated discharge values for the Athabasca Valles outflow channel ($1\text{--}2 \times 10^6$ m³/s) (Figure 6). These required melting areas equal and exceed the total area of the Central Elysium Planitia lava plains. While the low-end discharge estimate (1×10^6 m³/s) could be met by both peak and average melting rates over areas approximately equivalent to that of the Central Elysium Planitia lava plains (Figure 6), this would require complete collection and focusing of flow from the entire region to the head of Athabasca Valles. This condition is physically improbable as there exist many nested drainage basins within the $1\text{--}10 \times 10^6$ km² area of the Central Elysium Planitia volcanic plains which would direct meltwater flow along many different

pathways. Therefore, as has been found in previous investigations of outflow channel formation by lava-ice interactions (Cassanelli and Head, 2018), generation of the requisite discharge values requires melting to occur over unreasonably large areas, and by consequence, additional mechanisms to confine, accumulate, and release meltwater in outbursts at elevated discharge rates. It is worth noting, however, that the nominal discharge estimates utilized to evaluate Athabasca Valles are derived under bankfull flow assumptions, and are therefore likely to be substantial over-estimates of the actual discharge of the system (Wilson et al., 2004; Cassanelli and Head, 2018). Therefore, it remains possible for Athabasca Valles to have been carved at lower discharge rates with either a greater number of flooding events or through more sustained flow, scenarios that are explored in greater detail by Cassanelli and Head (2018) for a similar martian outflow channel system (Reull Vallis in Hesperia Planum). In addition, there exists a multitude of processes which can result in the confinement and release of meltwater (Cassanelli and Head, 2018) in bursts at greatly elevated discharge rates which can reach those peak estimates derived from the Athabasca Valles system. We later explore the mechanisms, discharge rates, and possible geomorphic signatures generated by meltwater confinement and release and the potential involvement of this process in a lava-ice origin of Athabasca Valles.

The final criterion established from the morphometry of the Athabasca Valles system requires increased incision at the channel head near the Cerberus Fossae source (Figure 2). To account for this geomorphic characteristic in a lava-ice interaction origin scenario, the majority of meltwater production would have to take place up-slope from the source region and to have focused melt directly to the channel source area. This is a difficult

requirement given that the apparent source of the channel is also the apparent source of the regions lava, which would then require the unreasonable flow of lava upslope (to generate meltwater upslope of the channel head) to areas where there are also little volcanic plains mapped (Figure 1). Therefore, due to this geomorphic characteristic, as well as the significant size of the system, the Athabasca Valles outflow channel poses the most stringent test of the lava-ice origin mechanism of any channel within Central Elysium Planitia. This difficulty, however, can be negated by the operation of meltwater confinement and release if sufficient quantities of water could be accumulated around the channel head area, a scenario we now assess.

7.3 Meltwater Confinement and Release

Evaluation of the melting rates generated by surficial lava-ice interactions for conditions relevant to the Central Elysium Planitia region indicates that, for melting over a reasonable source area scale, the peak discharge estimates for Athabasca Valles cannot be directly supplied (Figure 6). However, there are a multitude of mechanisms which could have occurred in conjunction with lava-ice melting to confine and release accumulated reservoirs of meltwater at greatly increased discharge rates. These mechanisms are reviewed by Cassanelli and Head (2018) and fall under three broad categories (relevant references and a more thorough discussion are provided by Cassanelli and Head, 2018).

(1) Meltwater confinement and accumulation within supraglacial or proglacial surface reservoirs created by snow, ice, or sediment/regolith damming with subsequent failure by (a) disruption of dams by melting, (b) increased water volumes and pressure, (c) drainage through ice sheet fractures, moulins, or permeable layers, (d) triggering from

external mechanisms such as landslides, avalanches, or ice mass failures delivering additional mass into the reservoir, or (e) increased pore pressures within the dam materials.

(2) Meltwater confinement and accumulation within englacial or subglacial reservoirs, with subsequent reservoir failure by drainage through ice sheet fractures, moulins, permeable layers, or subglacial drainage networks, or by the pressurization of subsurface reservoirs through meltwater addition causing floatation of the confining ice and rapid subglacial drainage, a primary mechanism for the generation of the Icelandic jökulhlaups (large glacial outburst floods) on Earth.

(3) Subglacial and englacial magmatic intrusion and melting with failure occurring by the same processes as previously discussed with the addition of the potential for volcanic melting to extend toward the ice cap margin, opening a pathway to the surface and discharging stored meltwater.

The confinement and release of meltwater by the mechanisms outlined above have been estimated to produce discharges as high as $10^7 \text{ m}^3/\text{s}$ in terrestrial flooding events, with the highest discharges typically associated with surface reservoir failure/overflow and jökulhlaup-like subglacial outbursts (O'Connor and Costa, 2004) (Figure 3c). These observed terrestrial discharge rates considerably exceed the peak discharge estimates for the Athabasca Valles and it is therefore plausible for meltwater confinement and release to have led to the formation of the outflow channel system. This plausible consistency is supported by the relatively rapid meltwater production rates involved specifically in lava-ice interactions (Cassanelli and Head, 2016, 2018) which would allow accumulation of large meltwater reservoirs without requiring sustained confinement conditions.

While terrestrial examples suggest that confinement and release mechanisms can generate adequately large discharge values to supply the Athabasca Valle channel system, it is valuable to assess the plausibility of this process under conditions specific to Central Elysium Planitia and our investigation. The most likely mechanisms for meltwater confinement and release in the surficial lava-ice interaction scenario we investigate, given the regional extent of the surface ice and the supraglacial lava flow emplacement geometry being assessed, are: (1) supra/englacial meltwater accumulation confined by an ice dam, with release triggered by flotation of the confining ice allowing drainage beneath the dam (Figure 3c) (much like the processes responsible for the generation of glacial lake outburst floods on Earth, such as the glacial Lake Agassiz events; Clarke et al., 2004), and (2) supra/englacial meltwater accumulation retained by the ice surface topography released by subsequent top-down supraglacial overflow (Smellie, 2006) leading to the erosion (by both mechanical and thermal processes; Marston, 1983; Russell et al., 2001) of a channel in the confining ice (similar to processes observed in some Icelandic jökulhlaup events; Russell et al., 2001; Smellie, 2006) (Figure 3c). Drainage of an accumulated meltwater reservoir by flotation and flow beneath a confining ice dam (Figure 3c) would require separation of the initially cold-based ice dam material from the subsurface. This could occur by hydro-fracturing or meltwater infiltration and melting at the base of the ice dam.

The discharge that can be produced by meltwater confinement and release is exponentially dependent upon the size of the outlet conduit through which water is released (O'Connor and Costa, 2004; O'Connor and Beebee, 2009). To assess the viability of the two most likely confinement and release mechanisms defined above, we

now evaluate the outflow conduit conditions required by each mechanism to reach the estimated discharge values of the Athabasca Valles system.

Ice dam flotation outflow:

To generate a simple estimate of the conduit dimensions required to meet the discharge values for the Athabasca Valles outflow channel by ice dam flotation outflow (Figure 3c), we assume the confinement and release process can be represented as an open reservoir draining through a conduit (the form and geometry of which we next attempt to constrain) along the base of the ice dam confining the reservoir, analogous to a free jet (the geometry of which is illustrated diagrammatically in Figure 3c). This reservoir and drainage geometry is commonly responsible for the generation of large glacial lake outburst floods and jökulhlaup events on Earth (Clarke et al., 1984; Waitt, 1985; Atwater, 1986; Costa and Schuster, 1988; Russell, 1989; Tweed and Russell, 1999; Tweed, 2000; Clarke et al., 2004; Björnsson, 2009).

Assuming the free jet geometry, the flow velocity at the outlet can be derived from the Bernoulli equation and is given by $\sqrt{2 g H}$ where H is the difference in height from the top of the reservoir to the mid-depth of the outlet conduit. Estimates of flow velocity by this method are oversimplified because steady flow is assumed and, more importantly, shear forces are neglected. While shear forces are not likely to have a significant effect on flow within the reservoir (at large distances from the confining walls), they will have great importance for the confined flow through the outlet conduit due to interactions with the conduit walls (particularly since the walls will be comprised of relatively rough geologic materials). The effect of shear forces, along with the likely turbulence of the conduit flow, will act to reduce the flow velocities and thus the discharge. Therefore, the

flow velocity estimates produced through this analysis will reflect maximum values and, consequently, yield minimum outlet conduit dimensions (since to reach any given discharge value, a maximum flow velocity will require a minimum flow area).

In order for meltwater release to occur by subglacial outflow as investigated here, the hydrostatic pressure at the base of the reservoir must equal or exceed the glaciostatic pressure at the base of the confining ice. At the maximum estimated regional ice thickness of 100 m, the onset of flotation would require an adjacent water reservoir ~91 m in depth. Assuming the maximum attainable melt reservoir depth of 100 m, limited by the maximum confining ice thickness, a conduit ~10 m in height could be created through confining ice flotation (by isostatic balance). After the onset of flotation the conduit size would vary in response to the balance between processes acting to enlarge and to close the conduit (Clarke et al., 2004) (which would lead to variations in the rate of discharge). In particular, thermal and mechanical erosion of the conduit walls during discharge would tend to increase the conduit size, but would be overcome by the glaciostatic pressure during the waning stages of flow as drainage of the reservoir proceeded and discharge diminished (Clarke et al., 2004). In our analysis, we assume a constant conduit height for simplicity but note this parameter, and by consequence the discharge, would vary with time throughout the course of outflow activity. Under the geometric conditions outlined above, the elevation difference from the reservoir surface to the conduit mid-depth, H , is ~100 m, which yields an outlet flow velocity of ~27 m/s. Assuming a rectangular outlet conduit cross-sectional area and a constant outlet flow velocity, an outlet conduit width of ~3.5-7.5 km is required to reach the $1-2 \times 10^6 \text{ m}^3/\text{s}$ discharge values estimated for Athabasca Valles. Given that Athabasca Valles heads along a ~75

km long section of the Cerberus Graben (Figure 2 and 7), the estimated peak discharges could be met even if the effects of turbulence and friction reduced the conduit outflow velocity by a factor of ~10.

Top-down supraglacial overflow:

Under the cold and dry ambient climate conditions which have dominated the Amazonian period on Mars (Madeleine et al., 2009, 2014), surface ice sheets are predicted to be capped by relatively thick layers of snow and firn (Cassanelli and Head, 2015). As a result, this porous layer of snow and firn could act to prevent the flotation of confining ice by accumulated meltwater (Smellie, 2006), preferentially leading to top-down supraglacial overflow (Figure 3c).

In order to estimate the size of the channel required to reach the necessary discharge by supraglacial overflow we utilize a version of the Manning equation for open-channel flow modified (Komar, 1979) to account for martian gravity:

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

where Q is discharge (m^3/s), W is channel width (m), d is channel depth (m), g_m is martian gravity, s is the sine of the down-flow slope of the channel base, R is the hydraulic radius (m), g_e is terrestrial gravity (9.81 m/s^2), and n is the Manning coefficient, which for ice has a value of $\sim 0.01 \text{ s/m}^{1/3}$ (Chu, 2014). In the initial stages of overflow, meltwater is likely to have been released as a widespread sheet flow which would be increasingly focused during continued outflow into a discrete channel (or many channels) as thermal and mechanical erosion of the confining ice progressed (Smellie, 2006). Assuming an initial channel slope approximately equivalent to that of the regional slope

(0.05°), we find that a sheet flow confined in an open channel 10 m in depth would need to extend to widths of ~12-24 km in order to supply the $1\text{-}2 \times 10^6 \text{ m}^3/\text{s}$ discharge estimate for Athabasca Valles. In later stages of overflow activity if flow were concentrated into a single channel with an increased slope (1°) and depth of 50 m, a channel width of only ~200-400 m would be required. These estimated overflow channel widths are well within the width constraints imposed by the ~75 km length of the Athabasca Valles channel head.

Our simplified analysis of the discharge rates generated by meltwater confinement and release (Figure 3c) indicate that the peak values estimated for the Athabasca Valles outflow channel could be provided for conditions relevant to Central Elysium Planitia. Could the requisite volumes of water also be supplied by these mechanisms? We earlier established that an approximate minimum melting area of 10^4 km^2 was required to provide the volumes of water necessary to carve the Athabasca Valles channel for a reasonable range of possible surface ice thicknesses. However, the disrupted terrain surrounding the Cerberus Fossae at the Athabasca Valles head does not extend over a comparable area (Figure 7). The disrupted terrain immediately surrounding the Cerberus Fossae structure at the head of the Athabasca Valles channel covers a total area of only ~1,700 km^2 (Figure 7). Our earlier results indicate that adequate volumes of water could not be produced by melting over this reduced area (Figure 5). These results, however, did not account for the possibility of flood recurrence, and were generated assuming the occurrence of only one melting and discharge event. If meltwater generation and accumulation occurred over a reduced melting area of 1,700 km^2 as mapped around the Cerberus Fossae (Figure 7), a total of ~2-9 melting and drainage events would be needed

to form the Athabasca Valles for a regional surface ice thickness of 100 m, while ~20-90 events would be required for a 10 m surface ice thickness.

As noted earlier, we find no strong evidence for melting over an expanded area of 10,000 km² around the Cerberus Fossae source region of the Athabasca Valles system. In principal, however, it is possible for melting over an area of this extent to have occurred for two reasons. (1) Individual lava flow units in Central Elysium Planitia exhibit typical areas of 10,000 km² (as described in Section 3) and could have been emplaced atop surrounding regional surface ice due to the exceptionally large effusion rates characteristic of the flood basalt volcanism of the area (up to 10⁶ m³/s; Keszthelyi et al., 2000, 2004). Some estimates suggest that individual lava flows in Central Elysium Planitia covering areas as large as 250,000 km² were emplaced over timescales as little as weeks (Jaeger et al., 2010), far more rapid than the melting timescales (Cassanelli and Head, 2016, 2018) associated with lava flows of the scale typical in Central Elysium Planitia (as described in Section 3). (2) Melting over this larger 10,000 km² area could have occurred and left little morphological signature due to obscuration by subsequent volcanic activity and lava flow emplacement which is known to have taken place in the system (Jaeger et al., 2007). Therefore, we consider melting over this larger area as a useful approximation for the upper limit of the possible melting area for any individual melting and discharge event. Over an area of 10,000 km², we find that only 1-2 melting and discharge events would be required for a regional surface ice thickness of 100 m, while 10-20 would be required for a 10 m surface ice thickness.

The estimated ranges in the total number of flooding events required to form Athabasca Valles (for melting areas of 1,700-10,000 km²) are generally consistent with

observations of the recurrence of glacial lake outburst floods and jökulhlaups on Earth (Baker and Bunker, 1985; Waitt, 1985; Benito and O'Connor, 2003). Multiple episodes of meltwater generation, confinement, and release could have been produced in Central Elysium Planitia by the re-accumulation of ice under ambient climate conditions (Head et al., 2006; Madeleine et al., 2009, 2014; Fastook et al., 2011) or during later obliquity excursions (Head et al., 2003a) interacting with recurrent volcanic activity in the region. This process requires multiple periods of volcanic activity spaced out in time with sufficient duration in between eruptions to allow the regional ice deposits to reform, a condition which is readily satisfied by the age distribution of the volcanic plains of the area (Vaucher et al., 2009a; Voigt and Hamilton, 2018). Following flooding activity, ablation of the regional surface ice and subsequent volcanic activity (as evidenced by the young lavas draping the Athabasca Valles channel; McEwen et al., 2012; Dundas and Keszthelyi, 2013) would be likely to leave minimal diagnostic geomorphic evidence of the confinement and release process.

7.4 Athabasca Valles Origin: Summary

In summary, we find that large-scale lava-ice interactions in Central Elysium Planitia could readily supply the water volumes necessary to form Athabasca Valles for reasonable thicknesses of surface ice within the region (Figure 5). While results indicate that lava-ice interactions are not able to directly supply meltwater discharges sufficient to source the outflow channel (Figure 6) a number of plausible processes could have led to the confinement and release of accumulated meltwater at greatly increased discharges. The rates of discharge that can be produced through the most likely confinement and release processes (englacial impoundment with failure by confining ice flotation and

supra/englacial accumulation released by supraglacial top-down overflow; Figure 3c) can reach the peak values estimated for Athabasca Valles and were found to be consistent with the geomorphic constraints of the channel system head. In addition, the extent of disrupted terrain along the Cerberus Fossae at the Athabasca Valles channel head suggests that meltwater confinement, accumulation, and outburst is not likely to have occurred over an area larger than $\sim 1,700 \text{ km}^2$ (Figure 7). At this area, anywhere from ~ 2 -100 melting, confinement, and outburst events would be required to supply the total volume of water needed to erode the Athabasca Valles channels. This required range of flooding events is consistent with the recurrence of glacial lake outburst floods and jökulhlaups on Earth and could have resulted from re-accumulation of ice under ambient climate conditions, or during later obliquity excursions, interacting with subsequent volcanic activity. Morphometric observations additionally have demonstrated that the Athabasca Valles outflow channel is most deeply incised near the channel source (Burr et al., 2002a; Keszthelyi et al., 2007). Meltwater confinement and release from the mapped $1,700 \text{ km}^2$ melting area (Figure 7) could conceivably have caused this erosional pattern, as the majority of meltwater production and release would have occurred at the channel head. Therefore, the formation of Athabasca Valles (and possibly other outflow channels in the region) could plausibly have been the result of surficial lava-ice interactions through meltwater confinement, accumulation, and release (Figure 3c).

8. Conclusions

The Central Elysium Planitia region of Mars is host to the most recent large-scale volcanic and fluvial activity on Mars. The region is characterized by extensive, geologically recent flood basalt volcanic plains incised by several outflow channel

systems. Recent improvements in the understanding of paleo-glaciation on Mars suggest that surface snow and ice accumulation was likely to have occurred in the Central Elysium Planitia during the Amazonian period. Furthermore, Central Elysium Planitia contains among the best evidence of past lava-ice interactions on Mars in the form of rootless cones which are preserved over wide areas of the region's volcanic plains. Therefore, Central Elysium Planitia is a region of great interest for understanding the thermal and hydrologic evolution of Mars into the Late Amazonian period.

Through a first-order geomorphic assessment we find that the presence of surface snow and ice and the operation of lava-ice interactions could have played a role in the formation of rootless cones through direct molten fuel-coolant interactions during advancement of flood lavas over regional surface ice/snow (Figure 3a). Surface snow and ice and the operation of lava-ice interactions may also have been involved in the formation of the platy-ridged texture of the Central Elysium Planitia volcanic plains by inducing divergent/convergent flow in the overriding lavas as a result of variations in ice surface elevation or melting rates (Figure 3b). While surface snow and ice and lava-ice interactions could have been involved in the formation of these features, they are not required. This is because rootless cones could have formed through the vaporization of atmospherically-emplaced ground-ice while platy-ridged terrain could have formed through variations in lava flow conditions without major influence of underlying snow and ice (due to the short timescales of lava emplacement relative to heat transfer and ice melting). In either case, the presence of surface snow and ice could have enabled more rapid lava flow emplacement through the formation of a lubricating meltwater lens generated by melting as documented in terrestrial experiments. This could help explain

the exceptional length of the Central Elysium Planitia lava flows which have previously been attributed to low lava viscosities, high effusion rates, and lava flow emplacement occurring largely through flow confined beneath a protective cooled upper crust. Thus, if surface snow and ice was directly involved in the formation of rootless cones and platy-ridged terrain, or indirectly by enabling greater lava flow emplacement rates, the extent of rootless cones and platy-ridged terrain could provide information on the age, thicknesses, and distribution of past glacial activity on Mars.

In addition to the potential geomorphic influences, surface snow and ice, and the operation of lava-ice interactions, could have led to the production of significant quantities of meltwater that may have contributed to the formation of the fluvial outflow channel features contained within the region. These potential contributions were evaluated through a case study of the Athabasca Valles outflow channel system, a well-characterized system which poses a challenging test of the lava ice origin scenario due to its scale, geomorphology, and proximity to the Cerberus Fossae source. Modeling of the top-down heating and melting induced by large-scale surficial lava-ice interactions using parameters and conditions appropriate to the Central Elysium Planitia region indicates that the volume of water needed to form the Athabasca Valles outflow channel (400-1,600 km³) could have been provided by melting over an area of $\sim 10^4$ to 10^6 km² while supply of the nominal discharge estimates ($1\text{-}2 \times 10^6$ m³/s) would require melting over an area of $\sim 10^6$ to $>10^7$ km² (Figure 5 and 6). The melting area range required to provide the necessary water volume overlaps the range between individual lava flow areas within Central Elysium Planitia and the total area of lavas within the region. Therefore, the water needed to carve the Athabasca Valles outflow channel could plausibly have been

provided by lava-ice interactions for reasonable surface ice thicknesses, particularly those approaching the maximum likely regional thickness of 100 m. In contrast, the melting areas required to provide the discharge necessary to form Athabasca Valles equal and exceed the total area of the Central Elysium Planitia lavas thereby requiring additional mechanisms to confine, accumulate, and release meltwater in outbursts at the required elevated discharge rates. We explored the most likely of the several possible meltwater confinement and release mechanisms (Figure 3c), (1) supra/englacial meltwater accumulation confined by an ice dam, with release triggered by flotation of the confining ice allowing drainage beneath the dam, and (2) supra/englacial meltwater accumulation retained by the ice surface topography released by subsequent top-down supraglacial overflow. Simple analysis of these processes indicates that the peak discharge values estimated for Athabasca Valles can be supplied within the geomorphic constraints of the outflow channel system. However, meltwater accumulation and release from the $\sim 1,700$ km² disrupted terrain at the Athabasca Valles channel head (Figure 7) would require anywhere from ~ 2 -100 melting, confinement, and outburst events to supply the total volume of water needed to erode channel. This required range of flooding events is consistent with the recurrence of glacial lake outburst floods and jökulhlaups on Earth and could have resulted from re-accumulation of ice under ambient climate conditions, or during later obliquity excursions, interacting with subsequent volcanic activity. Morphometric observations additionally have demonstrated that the Athabasca Valles outflow channel is most deeply incised near the channel source. Meltwater confinement and release from the mapped 1,700 km² melting source area (Figure 7) could plausibly have caused this erosional pattern, as the majority of meltwater production and release

would have occurred at the channel head. Therefore, the formation of the Athabasca Valles outflow channel system (and possibly other outflow channels in the region) by surficial lava-ice interactions through meltwater confinement, accumulation, and release offers a viable alternative to the problematic classical groundwater outburst models.

Acknowledgements

We are grateful for helpful reviews from John Smellie and an anonymous reviewer which improved the quality of this manuscript. Thanks are extended to Jay Dickson for assistance in generating CTX DEMs. We acknowledge support to JWH for participation in the Mars Express High-Resolution Stereo Camera Team (JPL 1237163).

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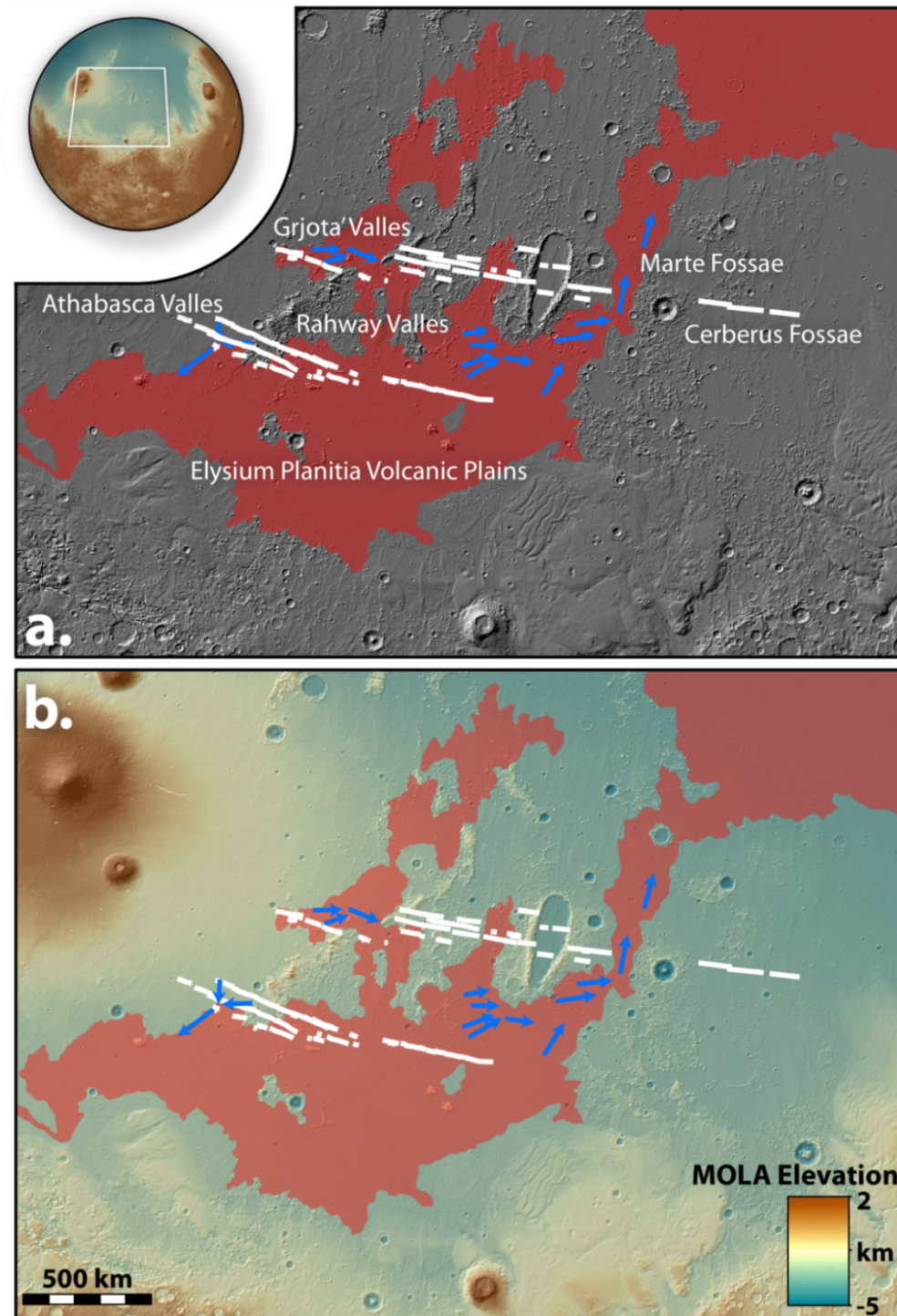


Figure 1. The Central Elysium Planitia region shown in Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001) hillshade data (a.) with overlain MOLA topography data (b.). The Late Amazonian smooth volcanic plains of the region (as mapped by Tanaka et al., 2014) are represented by the red shaded area, the fissures of the Cerberus Fossae system are mapped as white lines, and the flow paths of the region's major outflow channels are indicated by blue arrows (and labeled accordingly).

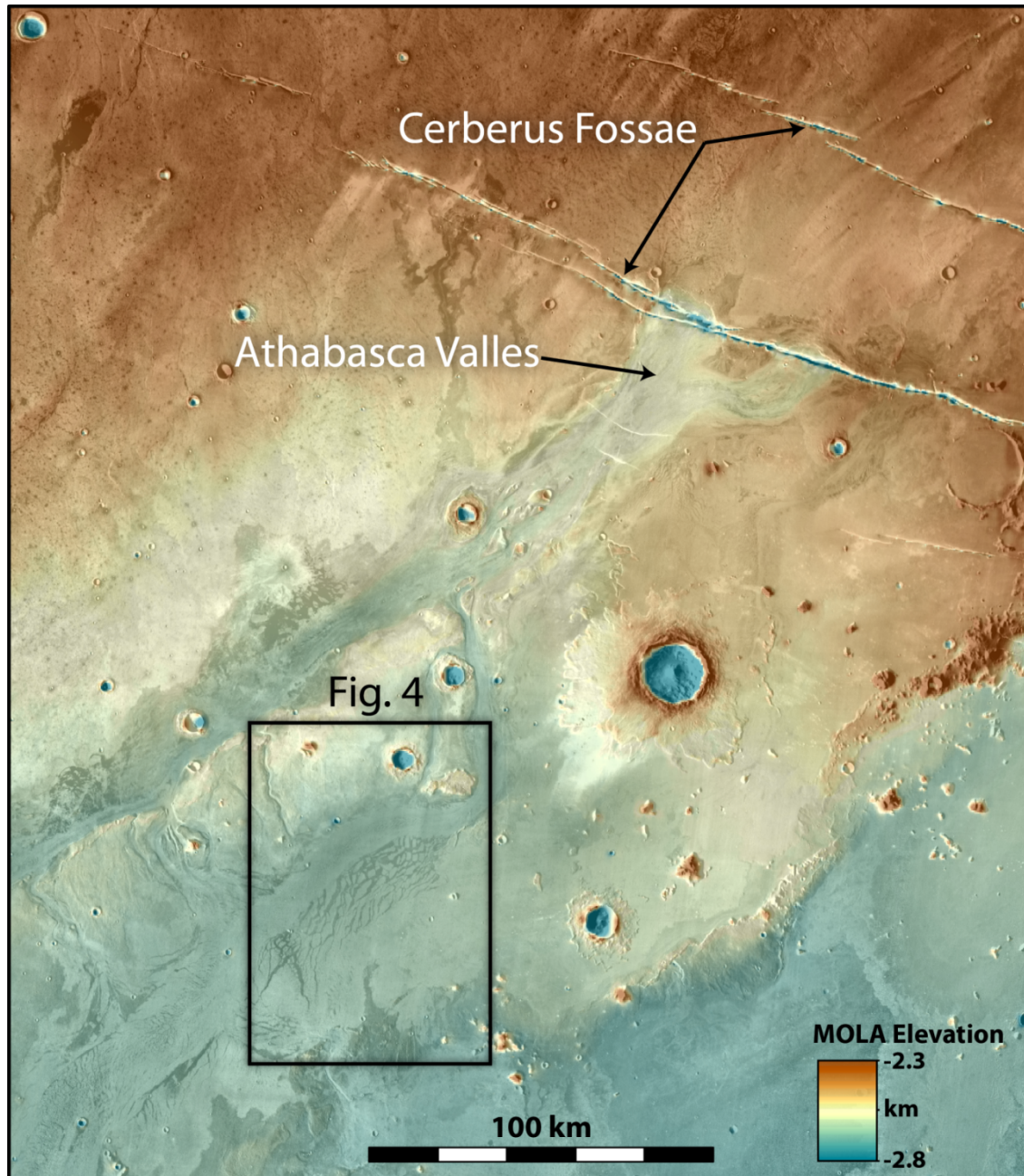


Figure 2. The head of the Athabasca Valles outflow channel at the Cerberus Fossae in Central Elysium Planitia shown in a mosaic of Thermal Emission Imaging System (THEMIS; Christensen et al., 2004) global IR data overlain by shaded MOLA topographic data.

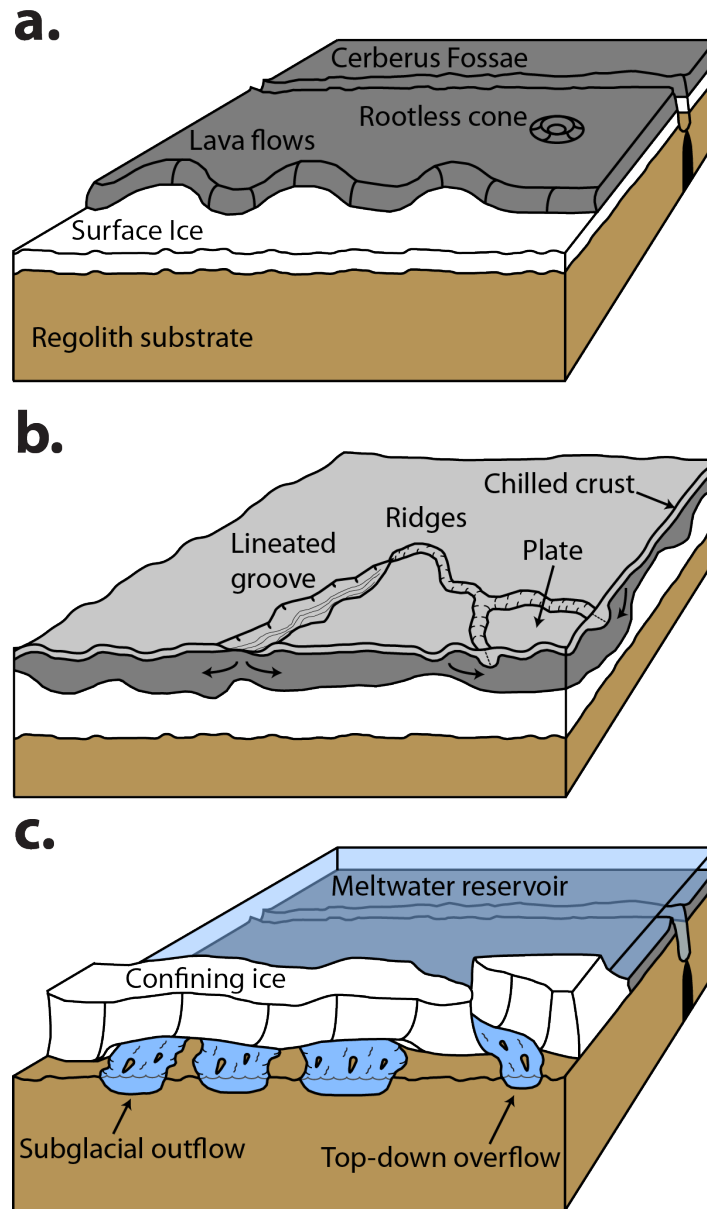


Figure 3. Conceptual synthesis of the involvement of large-scale surficial lava-ice interactions in the formation of (a.) rootless cones, (b.) platy-ridged terrain, and (c.) outflow channels in Central Elysium Planitia. Rootless cone formation is suggested to occur by direct molten fuel-coolant interactions during advancement of lava over regional surface ice/snow (discussed in Section 5). Platy-ridged terrain formation is proposed to occur by variations in ice surface elevation or melting rates resulting in divergent/convergent flow of the overriding lavas (discussed in Section 6). Outflow channel formation is proposed to occur by the generation of meltwater through lava-ice interactions followed by supra/englacial confinement and subsequent release of accumulated meltwater at elevated discharge levels (discussed in Section 7). Release of meltwater reservoirs (c.) could have occurred by (1) flotation of a confining ice dam allowing drainage beneath the dam, or (2) through top-down overflow eroding through the confining ice dam.

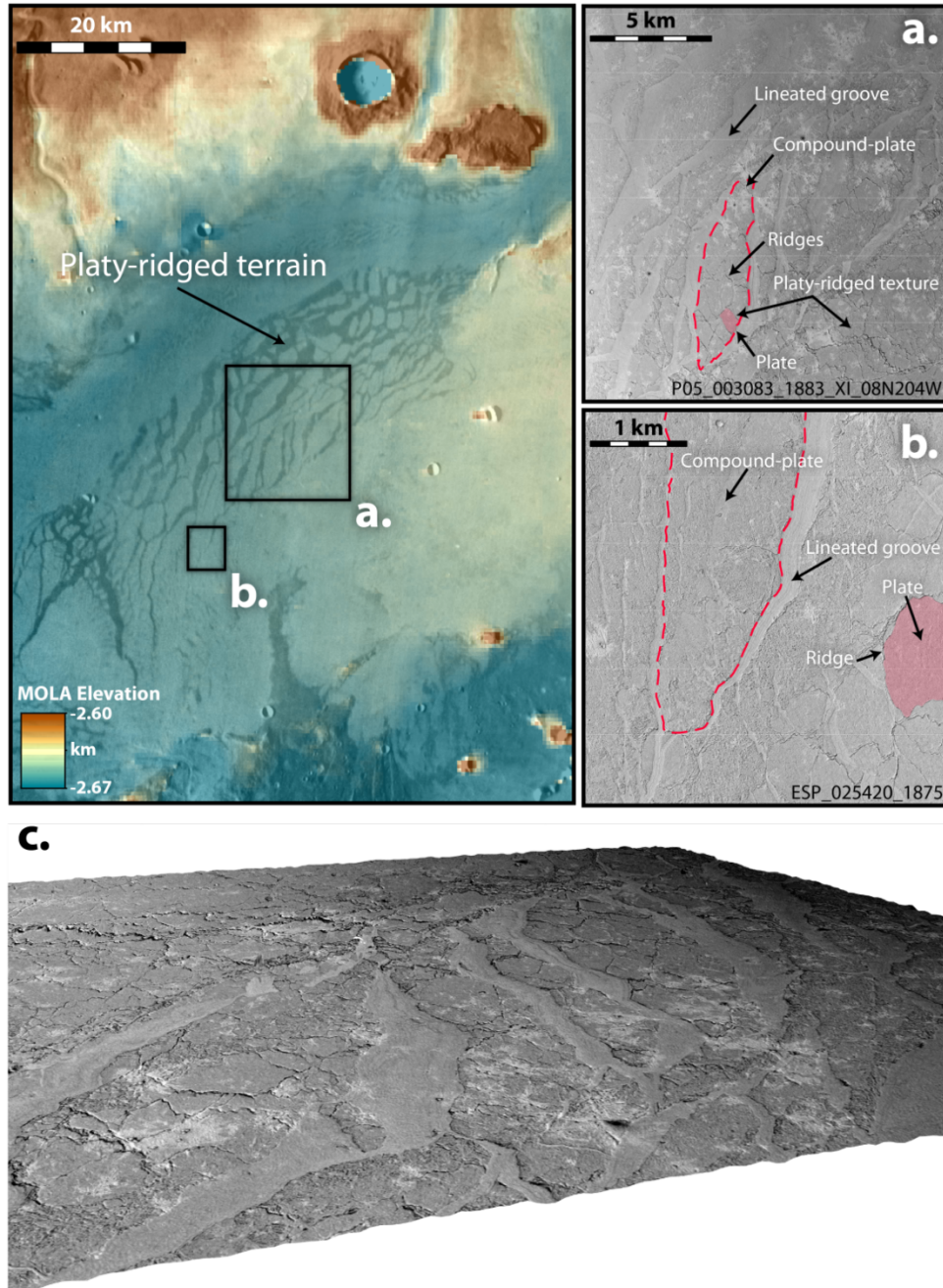


Figure 4. A region of the Central Elysium Planitia volcanic plains (context provided in Figure 2) exemplifying the platy-ridged terrain surface morphology shown in a mosaic of THEMIS data with overlain MOLA topography data. Higher resolution insets generated with (a.) Context Camera (CTX; Malin et al., 2007) (Image P05_003083_1883_XI_08N204W) and (b.) Hi Resolution Imaging Science Experiment (HiRISE; McEwen et al., 2007) (Image ESP_025420_1875) imagery illustrate the geomorphologic characteristics of the platy-ridge terrain in detail. (c.) CTX stereo pair digital elevation model (produced from Image P05_002938_1887_XI_08N204W and P05_003083_1883_XI_08N204W with 10X vertical exaggeration) perspective view of the platy-ridged volcanic plains surface morphology.

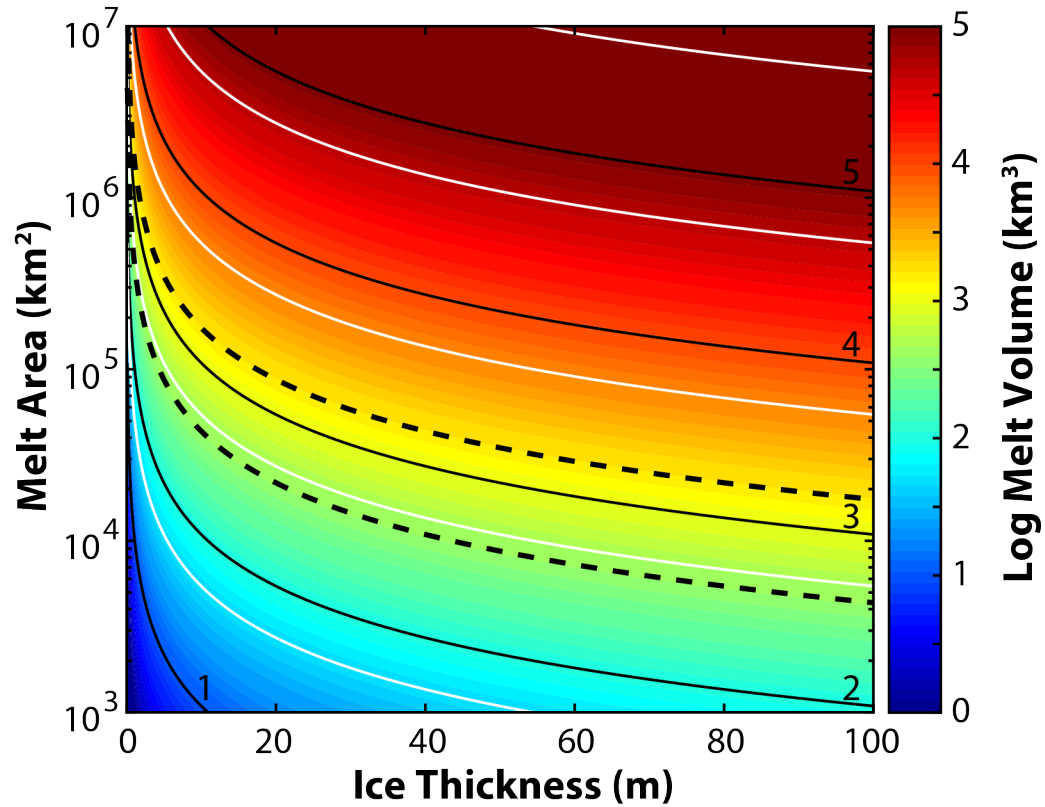


Figure 5. Contours of melt volume produced by lava-ice interactions as a function of the thickness of present regional surface snow and ice and the area of melting. The fine black lines represent integer contours of constant ($\log_{10}$) melt volume with the fine white lines plotted at 0.5 log units between the integer value contours. The nominal range of water volumes (400-1,600 km^3) estimated (Hanna and Phillips, 2006) to have been necessary to carve the Athabasca Valles outflow channel system are indicated by the bold dashed black lines.

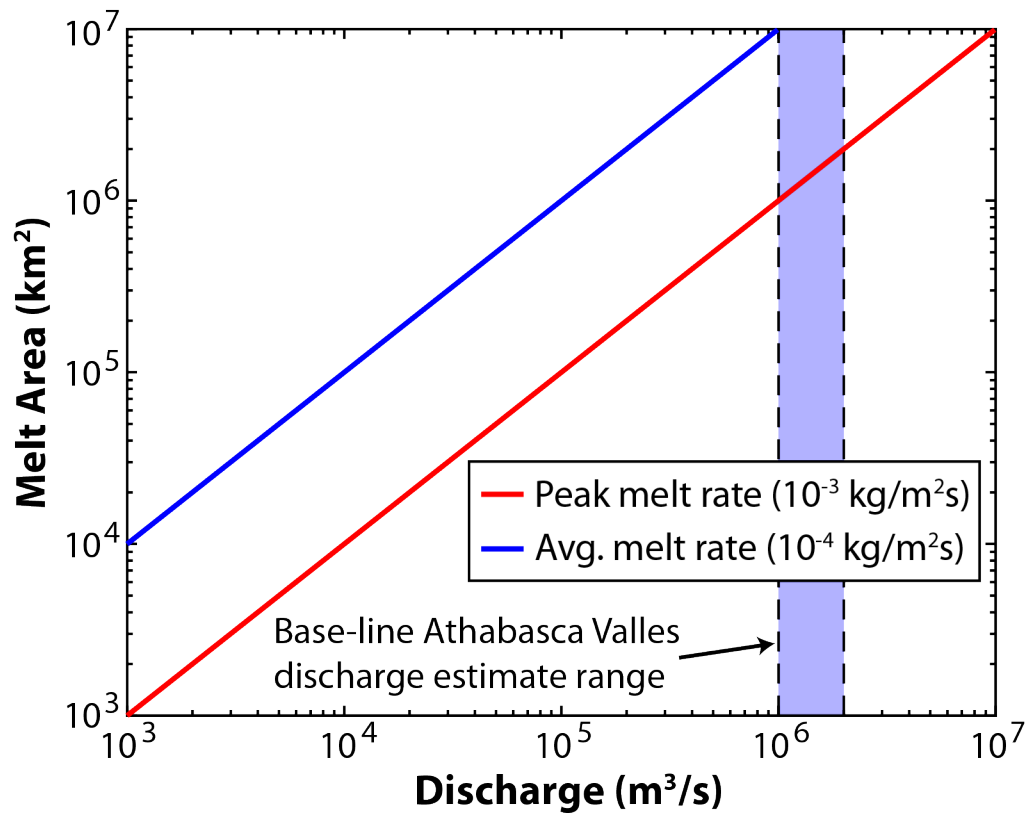


Figure 6. The total discharge produced by lava-ice interactions with parameters relevant to Central Elysium Planitia as a function of melting area. The nominal range of discharge estimates (Burr et al., 2002b) for the Athabasca Valles is shown in the blue shaded region for comparison.

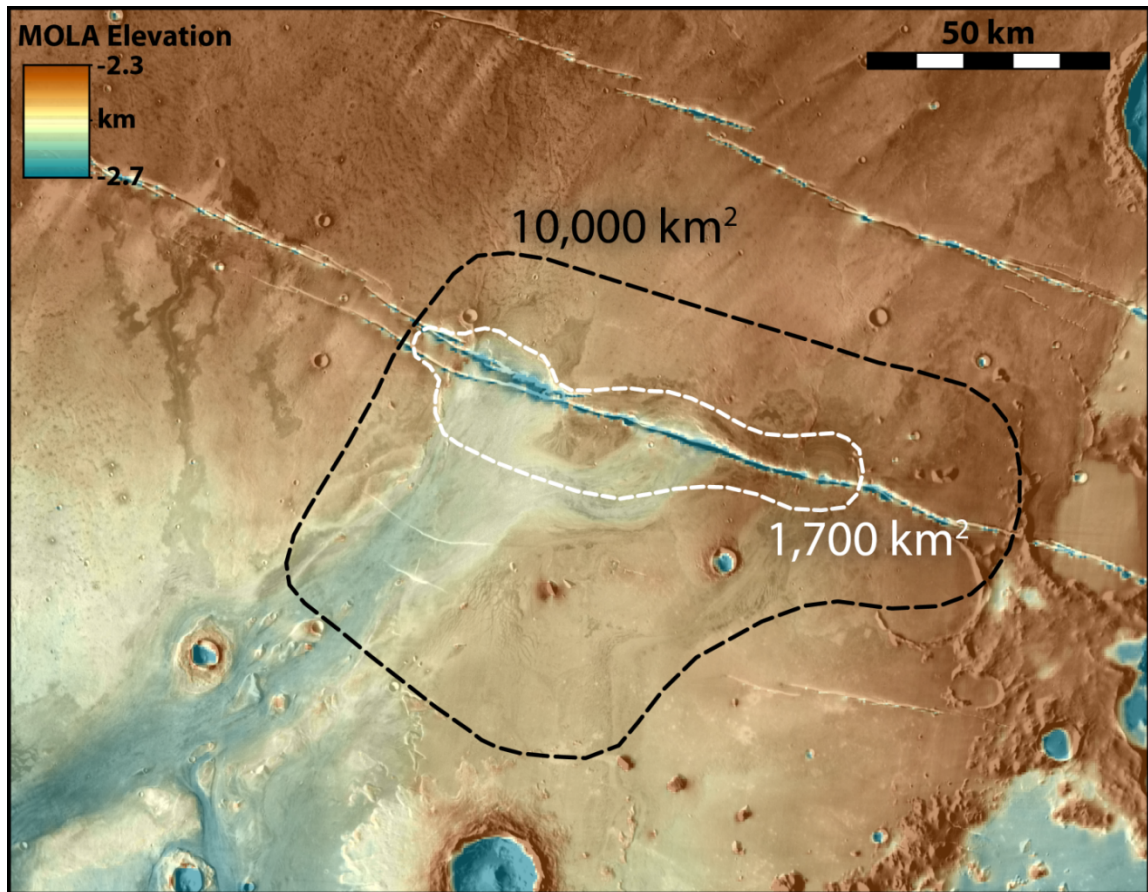


Figure 7. The head of the Athabasca Valles outflow channel along the Cerberus Fossae fissure shown in a global THEMIS mosaic with overlain shaded MOLA elevation data. An area of 10^4 km^2 is mapped for scale reference along with the disrupted terrain around the Cerberus Fossae covering $\sim 1,700 \text{ km}^2$.

Chapter 4: Glaciovolcanism in the Tharsis Volcanic Province of Mars:
Implications for Regional Geology and Hydrology

James P. Cassanelli^a & James W. Head^a

^aDepartment of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

Manuscript under review in: *Planetary and Space Science*, Manuscript Ref:

PSS_2018_235

Abstract

The Tharsis region of Mars is a vast volcanic plateau which hosts the immense Tharsis Montes shield volcanoes. The Tharsis region is suggested by multiple lines of morphologic and modeling evidence to have been a site of coincident volcanic and glacial activity throughout the majority of the geologic history of Mars. The prolonged and overlapping histories of volcanism and glaciation within the Tharsis region raise the likelihood of widespread surficial glaciovolcanism, a possibility which is supported by the recognition of glaciovolcanic landforms in the region by past investigations. Given this likelihood, we perform an exploratory study to assess the potential role that surficial glaciovolcanism may have played in the geologic and hydrologic history of the Tharsis region. We first review the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, in order to outline relevant conditions and parameters. The outlined volcanic and glacial conditions are then used in conjunction with results and predictions from past modeling of surficial glaciovolcanic processes to assess the potential role of glaciovolcanism in the formation of the Tharsis region's major tectonic and hydrologic features. Specifically, we assess glaciovolcanic formation models for the aureole apron units which surround the Olympus Mons shield volcano, the possible Syria-Thaumasia block/Solis Planum mega-thrust, the valley networks contained within the highland terrains and incised into the flanks of the shield volcanoes in the region, and finally, the large outflow channels.

We find the origin of the Olympus mons aureole apron units can plausibly be explained by the emplacement of lava flows emanating from the edifice atop surrounding

regional surface ice. An origin by this mechanism can account for the great distances over which the apron units are deposited through enhanced supraglacial lava flow emplacement velocities with apron ridges and distal ramparts produced by post-emplacement subsidence. The potential glaciovolcanic origin of the Syria-Thaumasia block/Solis Planum mega-thrust was investigated through a formation model involving detachment and deformation along a weak, volcanically-buried ice layer produced by the accumulation of superposed volcanic material. Due to the thermal disequilibrium of ice buried to the estimated depths of detachment for the Syria-Thaumasia block/Solis Planum mega-thrust, formation by this mechanism requires exceptionally rapid accumulation of large thicknesses of volcanic material and is therefore not favored. We next examined a possible glaciovolcanic origin of the regional highland valley networks within the Tharsis region by the dewatering of crust and thawed ice uplifted to the surface by a Solis Planum mega-thrust. We find that, while the position and distribution of the regional valley network systems can be accounted for by this mechanism, the geomorphology of the systems is inconsistent with formation in this manner, and thus previously developed origin models are favored. To explore the possible surficial glaciovolcanic formation of the volcanic valley networks of the Tharsis region, we performed a case study of the Ceraunius Tholus valley network system. We find that an origin by surficial glaciovolcanism better accounts for the formation of these systems than prior ice sheet basal melting models by producing greatly elevated meltwater production rates and consistent morphologic characteristics. In our final analysis, we explored possible surficial glaciovolcanic origins for the fossae/fracture-sourced and basin/chaos-sourced outflow channels of the Tharsis region through case studies of the Mangala Valles and

Kasei Valles systems, respectively. We find that the formation of the Mangala Valles system by supra-ice emplacement of Mangala Fossa lava flows, with subsequent meltwater accumulation and release is able to supply the water volumes and discharges needed to erode the channel while conforming to the geomorphic constraints of the system. Finally, we find that the formation of the Kasei Valles system can be plausibly explained by recurring surficial glaciovolcanism within the channel's drainage basin producing meltwater and runoff subsequently collected and released from reservoirs within Echus Chasma and the north-trending portion of the channel system. This formation mechanism is a natural consequence of the extensive, sustained volcanism and glaciation in the region and is able to supply the water volumes and discharges required to form the channel while also accounting for the distributed and recurring inputs inferred to have contributed to the system. Ultimately, we conclude that surficial glaciovolcanism may plausibly have contributed to the formation of many of the tectonic and fluvial features in the Tharsis region, offering advantages over prior formation models, particularly for the large basin/chaos-sourced outflow channels concentrated in the area.

1. Introduction

The Tharsis rise of Mars (Figure 1a) is the most prominent volcanic province on the planet and arguably within the solar system. The Tharsis region also represents the most active center of volcanism on the planet, with a history of volcanic activity spanning from the Noachian to the very recent Late Amazonian period (discussed in detail in the following section). Rising to a mean elevation of ~ 2.3 km (based on our defined extent of the Tharsis region, later detailed), and with localized elevations in excess of 20 km (Figure 1a), the Tharsis rise is also a region predicted (Forget et al., 2013; Wordsworth et

al., 2013, 2015; Wordsworth, 2016) to exhibit mean annual surface temperatures far below the global average during the early history of Mars. This prediction is a result of the influence of adiabatic cooling in a thicker early CO₂ atmosphere (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) causing substantially reduced temperatures at the elevations of the Tharsis region. As a result of the depressed mean annual surface temperatures in the presence of a thicker atmosphere, the Tharsis rise is predicted to act as a cold trap (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016), leading to the accumulation of regional snow and ice above a +1 km equilibrium line altitude (Head and Marchant, 2014) within the region under both “cold and icy” (global mean annual surface temperature ~225 K; (Wordsworth et al., 2015) and “warm and wet” (global mean annual surface temperature >273 K; Wordsworth et al., 2015) end-member climate scenarios (the “icy highlands” climate scenario) (Figure 1b). Beyond the early Noachian and Hesperian history of Mars, following the loss of the thicker atmosphere (Carr and Head, 2010; Lammer et al., 2013; Jakosky et al., 2017) and thus adiabatic cooling conditions, abundant evidence (later outlined in detail) suggests regional accumulation of snow and ice has continued throughout the Amazonian period within the Tharsis rise region.

Therefore, the Tharsis region is a site where coincident volcanic and glacial activity have occurred throughout the majority of the history of Mars, raising the possibility for widespread and sustained glaciovolcanism. Here we perform an exploratory study to assess the potential role of glaciovolcanism (Allen, 1979; Squyres et al., 1987; Smellie and Chapman, 2002; Head and Wilson, 2007; L. Wilson and Head, 2007; Wilson and Head, 2009; Edwards et al., 2013, 2014; Wilson et al., 2013; Scanlon et al., 2014; Smellie

and Edwards, 2016) in the geologic and hydrologic history of the Tharsis region. To perform this assessment, we first review the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, in order to constrain relevant conditions and parameters. We then evaluate the potential role of glaciovolcanism in the formation of major tectonic and hydrologic features within the region including, volcanic aureole apron units, the Solis-Planum mega-thrust, valley networks, and outflow channels (Figure 1). Analysis of the role of glaciovolcanism in the formation of these features is carried out by first outlining formation criteria through reviews of the geomorphology of each feature. The outlined formation criteria of each feature are then evaluated for consistency against predictions from possible genetic surficial glaciovolcanic models informed by previous modeling work (Cassanelli and Head, 2016, 2018a).

2. Tharsis Region Volcanism

The nature of surficial glaciovolcanic activity is controlled in large part by the rate and volume of lava emplacement (*e.g.* Squyres et al., 1987; Wilson and Head, 2002a; L. Wilson and Head, 2007; Smellie and Chapman, 2002; Head and Wilson, 2007; Wilson et al., 2013; Edwards et al., 2014; Cassanelli and Head, 2016, 2018a; Smellie and Edwards, 2016), and thus by the prevailing style of volcanism. Therefore, in order to evaluate the nature of possible surficial glaciovolcanic processes in the Tharsis region, we first outline the styles and characteristics of volcanism which have occurred throughout the history of the area.

The Tharsis rise is a vast volcanic plateau covering a total area of $\sim 3 \times 10^7$ km² (nearly 25% of the surface of Mars) (Phillips et al., 2001; Tanaka et al., 2014), comprised of an

estimated $\sim 3 \times 10^8 \text{ km}^3$ igneous material (Phillips et al., 2001), and containing 13 large, distinct volcanic edifices (Tanaka et al., 2014) (Figure 1a). Here we define the extent of the Tharsis region through the aggregation of mapped units (Tanaka et al., 2014) on the basis of volcanic origin, unit elevation, and unit surface morphology. As a result, we therefore include Olympus Mons, its surrounding apron unit, and Lunae Planum in our definition of the Tharsis region, yielding a total defined area of $\sim 2.8 \times 10^7 \text{ km}^2$ (Figure 1) (within the range of extents previously defined for the Tharsis region). While there are no firm constraints on the inception of the accumulation of the Tharsis rise, the feature is canonically interpreted to have been largely in-place by the end of the Noachian period $\sim 3.7 \text{ Ga}$ (Phillips et al., 2001; Carr and Head, 2010) (although the bulk of Tharsis formation may have occurred later; Bouley et al., 2016; Citron et al., 2018). Following the Noachian, volcanic activity in the Tharsis region continued at a reduced rate throughout the Hesperian and into the Amazonian periods (Tanaka et al., 1988, 2014; Werner, 2009), to as recently as the last several 1-10's of Ma (Hartmann et al., 1999; Neukum et al., 2004; Hauber E. et al., 2011; Mangold et al., 2010; Richardson et al., 2017). Therefore, volcanic activity within the Tharsis region has been sustained throughout most of the geologic history of Mars.

The Tharsis rise is interpreted to have formed through hot-spot-like volcanism above a mantle upwelling, resulting from either predominantly crustal uplift (Carr, 1974; Greeley and Spudis, 1981; Phillips et al., 1990) or volcanic construction by the accumulation of lavas (Solomon and Head, 1982). The uplift origin of the Tharsis rise is supported (Carr, 1974; Plescia and Saunders, 1980, 1982; Phillips et al., 1990) by the broad domical shape of the feature (Figure 1a), the presence of ancient cratered terrains at

high elevations within the region, and extensional fractures in the region exhibiting generally radial trends. In an uplift origin scenario (Carr, 1974; Plescia and Saunders, 1980), the majority of the Tharsis topography is attributed to lithospheric uplift, with the surface later veneered by deposition of relatively thin lavas through plains volcanism, accompanied by the construction of the massive shield volcanoes. However, the uplift origin is challenged by disagreement between the orientation of tectonic features in the region and stresses predicted for lithospheric uplift of Tharsis, and by unreasonable thermal and chemical anomaly requirements for mantle support (Solomon and Head, 1982). In a constructional origin scenario (Solomon and Head, 1982), the Tharsis topography is attributed chiefly to the accumulation of volcanic deposits, with a substantial additional component of isostatically-compensated deposits residing at depth.

Here, we are interested in assessing the operation of glaciovolcanic processes in the Tharsis region, and the contributions they may have had toward the regional geology. Therefore, we adopt the constructional origin interpretation of Tharsis rise (Solomon and Head, 1982) as the greater extent of extrusive volcanic activity in this scenario would result in increased surficial glaciovolcanic activity. We note that, if instead Tharsis originated primarily by uplift, the amount of glaciovolcanic activity that would have taken place in the region would be consequently reduced (with the possibility of increased subsurface glaciovolcanism generated by increased intrusive magmatism; Phillips et al., 1990).

The Tharsis region as we have defined (Figure 1a) encompasses a range of geologic units (Tanaka et al., 2014) emplaced in a range of volcanic styles throughout the history of the Tharsis region including shield, plains, and flood basalt volcanism (Solomon and

Head, 1982) (Figure 2a). The volcanic edifices within the region and the main constituents of the Tharsis bulge itself (the Tharsis Montes and the adjacent plateau plains including Daedalia, Sinai, Solis, Syria, and Thaumasia Planum) are interpreted to have been formed through a combination of shield and plains volcanism (Figure 2a). Volcanic units emplaced in a flood basalt mode, inferred by their association with the Hesperian ridged plains volcanic units (Head et al., 2002) as evidenced by their surface ages and shared wrinkle-ridge morphology (Solomon and Head, 1982), are contained predominantly in the eastern portions of the Tharsis region (Figure 2a) (exemplified by Lunae Planum). Here the extent of surface units in the Tharsis region emplaced by flood basalt style volcanism have been defined through the aggregation of mapped (Tanaka et al., 2014) volcanic units in the region (Figure 1) which exhibit wrinkle-ridge surface morphology (Figure 2a). Explosive volcanism is an additional style of volcanic activity expected to have taken place within the Tharsis region during its course of development due to the prevalence of low atmospheric pressures (*e.g.* Mouginis-Mark et al., 1982; Edgett, 1997; Head and Wilson, 1998; Wilson et al., 1998; Hynek et al., 2003; Lionel Wilson and Head, 2007; Wilson and Head, 2009; Kerber et al., 2012). Past analyses (Wilson and Head, 2009) indicate that the supraglacial deposition of both heated and cooled volcanic tephra generated by explosive eruptions acts almost exclusively to protect the ice against ablation. Therefore, we have not directly considered this process in our analysis of surficial glaciovolcanic interactions, though we note, where relevant, the potential important implications of ice preservation by tephra burial.

The history of the emplacement of these various volcanic units within the Tharsis region (Solomon and Head, 1982) likely began in the Noachian period with intense,

extensive shield and plains volcanism constructing the bulk of the rise. Following the early phase of Tharsis construction in the Noachian period, shield and plains volcanism in the region continued at a diminished rate throughout the Hesperian with the additional emplacement of the extensive flood basalt volcanic plains in the eastern portions of the region (Figure 2a). Volcanic activity in the Tharsis region then continued through the most recent Amazonian period of Mars history (spanning the last >3 Gyr) through further plains volcanism and development of the large shield constructs of the region. This emplacement history and the distribution and timing of volcanism within the Tharsis region is illustrated by the mapped (Tanaka et al., 2014) current surface ages of the region's volcanic units (Figure 2b) as determined through stratigraphic and crater age dating. The mapped surface ages (Figure 2b) indicate that the majority of the Tharsis region is covered by plains volcanic units dating from the Hesperian to Amazonian (Figure 2b) period. The oldest volcanic units exposed at the present surface reside predominantly in the eastern and southeastern portions of the Tharsis rise and are associated with volcanic and flood basalt plains. The most recent volcanic activity in the Tharsis region is associated with the large shield volcanic constructs of the area which have been resurfaced by volcanic deposits into the Late Amazonian (Figure 2b) and which have remained centers of continued volcanic activity to as recently as the last 1-10's of Ma (Hartmann et al., 1999; Neukum et al., 2004; Hauber E. et al., 2011; Mangold et al., 2010; Richardson et al., 2017).

Critical parameters required for our analysis of surficial glaciovolcanic processes include the total thickness and area of lava accumulation as well as the thicknesses and areas of individual lava flows. Under the constructional Tharsis origin interpretation we

have adopted, the total accumulated thicknesses of lava resulting from the operations of shield and plains volcanism in the region are on the order of ~1-10 km (*i.e.* the approximate range of total Tharsis topography) (Solomon and Head, 1982). The total thickness of the flood basalt volcanic plains located on the eastern side of the Tharsis region (Figure 2a) are estimated at ~1-3.5 km (Plescia and Saunders, 1980; De Hon, 1982; Watters, 1991) and cover a total area of $\sim 4.5 \times 10^6 \text{ km}^2$ (Figure 2a). Measurements and estimates of the thicknesses and areas of individual lava flows produced by shield volcanism in the Tharsis region yield typical values on the order of ~50 m and $\sim 500 \text{ km}^2$, ranging from ~1-220 m and $\sim 2-25,000 \text{ km}^2$ (Baloga et al., 2003; Warner and Gregg, 2003; Basilevsky et al., 2005; Hiesinger et al., 2007; Mouginis-Mark and Rowland, 2008; Richardson et al., 2017). Typical thicknesses and areas of individual lava flows generated by plains volcanism in the Tharsis region, and in other similar deposits on Mars (chiefly Elysium Planitia which contains the among the youngest and best preserved volcanic plains deposits on Mars; (Plescia, 2003; Keszthelyi et al., 2004; Vaucher et al., 2009a), average to ~20 m and $\sim 10,000 \text{ km}^2$, and range from ~1-200 m (Scott and Tanaka, 1980; Simon et al., 2014) and $\sim 1,000-2.5 \times 10^5 \text{ km}^2$ (Scott and Tanaka, 1980; Plescia, 1990; Sakimoto and Gregg, 2001; Berman and Hartmann, 2002; Fuller and Head, 2002; Keszthelyi et al., 2000, 2004; Plescia, 2003; Werner et al., 2003; Lanagan, 2004; Vaucher et al., 2009b, 2009a; Jaeger et al., 2010; Simon et al., 2014).

Summary: In summary, the volcanic history of the Tharsis region is characterized by widespread and intense shield and plains volcanism in the Noachian period during which time the bulk of the rise was constructed. The intensity of volcanic activity in the area generally declined into the Hesperian with the exception of the extensive flood basalt

volcanic plains which were emplaced in the eastern portions of the region during this period. Volcanic activity within the Tharsis region continued throughout the Amazonian period at a diminished rate and became increasingly focused toward the development of the major shield constructs of the region. The characteristics of each of the different volcanic styles which have operated in the Tharsis region, including the timeframe of activity, the areas and thicknesses of total lava accumulation, and the areas and thicknesses of individual lava flows, are summarized in Table 1.

Having outlined the nature and characteristics of the volcanic history of the Tharsis region we now turn to a review of the glacial history of the region.

3. Tharsis Region Glaciation

Due to the significant elevation of the Tharsis region (Figure 1a), the majority of which is generally agreed to have formed by the end of the Noachian period (Phillips et al., 2001; Carr and Head, 2010), models of the early Mars climate (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) predict Tharsis to have been a location of regional snow and ice accumulation during the Noachian and early Hesperian periods. This is because, in the presence of a thicker early Mars CO₂ atmosphere, adiabatic cooling results in substantially decreased mean annual surface temperatures at the elevations of the Tharsis region causing Tharsis to act as a cold trap (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). This is a robust prediction found to occur not only in models predicting a cold and icy climate (where the global mean annual surface temperature is ~225 K; (Wordsworth et al., 2015), but also in forced warm and wet end-member climate models where the global mean annual surface temperature is raised above 273 K through the addition of a uniformly absorbing grey gas

(Wordsworth et al., 2015). Under these predicted early Mars climate conditions, snow and ice become sequestered within the Tharsis region due to a lack of ablation, thus leading to the accumulation of snow and ice deposits to thicknesses which are limited by the available global supply of surface/near-surface water. The current available surface/near-surface water inventory on Mars is equivalent to a ~34 m global equivalent layer (GEL) of ice (Carr and Head, 2015). Plausible ranges for the early Mars, Late Noachian/Early Hesperian available surface/near-surface water inventory range from ~1-5X the current available global surface water inventory, yielding a range of ~34-170 m GEL (Fastook and Head, 2015). However, recent evaluation (Carr and Head, 2015) of the available global surface/near-surface water supply during these early periods of Mars history suggests a total global available inventory not significantly different from the present (~34 m GEL). Within these global available surface/near-surface water inventory constraints, climate (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) and glacial (Fastook and Head, 2015) modeling predict the accumulation of regional ice sheets above a +1 km equilibrium line altitude (ELA) (Figure 1b) to average thicknesses of ~278-725 m, with maximum ice sheet thicknesses reaching ~0.7-2.8 km within deep craters.

During the later periods of Mars history, beyond the Late Noachian to Early Hesperian, the mass of the martian atmosphere declined by escape and loss processes resulting in a decrease of the surface pressure from ~1 bar toward that of the present day, ~6 mbar (Carr and Head, 2010; Lammer et al., 2013; Jakosky et al., 2017). This loss in atmospheric mass and surface pressure would have been accompanied by a decay in the strength of the adiabatic cooling effect, leading to a mean annual surface temperature

distribution governed by radiative equilibrium (Wordsworth et al., 2013). Consequently, the distribution of mean annual surface temperatures would be controlled predominantly by latitudinal position, with minimal influence from elevation. Additionally, the loss of the insulating greenhouse effect produced by the thicker CO₂ atmosphere would lead to globally reduced mean surface temperatures. Climate models of the present day atmospheric conditions of Mars (Madeleine et al., 2014), with a 6 mbar CO₂ atmosphere, indicate a mean annual global surface temperature of ~210 K, a substantial reduction from the ~225 K which is sustained under the early thicker 1 bar CO₂ atmosphere (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016).

Under these later atmospheric conditions, the highly elevated Tharsis region would no longer act as a cold trap, ceasing the preferential regional accumulation and sequestration of snow and ice. However, there are several lines of evidence which suggest that more localized accumulation of snow and ice continued throughout later Mars history within the Tharsis rise region. Evidence supporting this continued accumulation of snow and ice includes: (1) Climate modeling studies predicting the accumulation of ice within the Tharsis rise region (Forget et al., 2006; Madeleine et al., 2009, 2014). (2) The presence of extensive fan-shaped deposits superposed upon Amazonian-aged lavas on the western flanks of the Tharsis Montes which are inferred to have been produced through cold-based tropical mountain glaciers (Head and Marchant, 2003; Shean et al., 2005; Forget et al., 2006; Fastook et al., 2008; Kadish et al., 2008; Head et al., 2010). (3) The presence of ring-mold craters on the Tharsis Montes interpreted to form due to impacts into an ice-rich substrate (Head and Weiss, 2014). (4) The distribution of ice-bearing latitude-dependent mantling deposits to mid-to-low latitudes (Kreslavsky and Head, 2002; Head

et al., 2003a, 2010; Madeleine et al., 2009; Schon et al., 2009; Fastook and Head, 2014; Dundas et al., 2018). (5) The preservation of ice-cored pedestal craters within the Tharsis region and at comparable latitudinal bands which are interpreted to form through ejecta armoring of surface snow and ice (Kadish et al., 2010; Kadish and Head, 2011; Schon and Head, 2012; Fastook and Head, 2014). (6) The widespread presence of layered ejecta craters interpreted to have required surface snow and ice to form (Weiss and Head, 2013, 2014, 2017). (7) Deposits of ice-bearing lineated valley fill, lobate debris aprons, and concentric crater fill (Head et al., 2006; Madeleine et al., 2009; Levy et al., 2010; Fastook et al., 2011; Fastook and Head, 2014; Levy et al., 2014). (8) The detection of abundant excess ground-ice contained within the nearby Arcadia Planitia through Mars Reconnaissance Orbiter Shallow Radar (SHARAD; Seu et al., 2004) instrument measurements (Bramson et al., 2015).

These lines of evidence can be divided into two broad categories. (1) Evidence for the formation of cold-based tropical mountain glaciers on the Tharsis Montes, indicative of more concentrated/localized snow and ice accumulation (items 1-3). The formation of cold-based tropical mountain glaciers on the western flanks of the Tharsis Montes is predicted to result from preferential snow and ice deposition due to westerly atmospheric circulation and orographic upwelling (Forget et al., 2006; Fastook et al., 2008). The distribution of the fan-shaped deposits on the western flanks of the Tharsis Montes suggest that cold-based glaciers covered an area of at least $\sim 1.6 \times 10^5 \text{ km}^2$, with predictions from glacial modeling (Fastook et al., 2008) indicating total areas as large as $\sim 5 \times 10^5 \text{ km}^2$ with ice thicknesses ranging from ~ 100 -4,000 m, averaging $\sim 2,000$ m. (2) Evidence provided by widespread regional features, such as the pedestal craters,

indicating a more distributed/regional pattern of snow and ice accumulation (items 4-8). This class of features is inferred to result from the redistribution of volatiles from the polar regions toward mid-to-low latitudes during periods of increased planetary obliquity (Kreslavsky and Head, 2002; Head and Marchant, 2003; Head et al., 2003a, 2005; Forget et al., 2006; Madeleine et al., 2009, 2014). While these features indicate the occurrence of distributed snow and ice accumulation, they do not provide precise constraints on the total areal distribution. The average thicknesses of surface ice indicated by these features is ~50 m, ranging from ~1-100 m with the lower bound corresponding to typical thicknesses of latitude dependent mantle deposits (Kreslavsky and Head, 2002; Schon et al., 2009) and the upper bound to maximum estimates of surface ice thickness derived from pedestal craters (Kadish et al., 2010).

Summary: In summary, the glacial history of the Tharsis region is characterized by the accumulation of widespread, regional ice sheets during the Noachian and Early Hesperian periods, transitioning to more focused/localized glaciation during the Hesperian and throughout the Amazonian periods. The timeframe, general distribution, and ice thicknesses associated with these various forms of glacial activity occurring throughout the history of the Tharsis region are summarized in Table 2.

With the history and nature of both volcanic and glacial activity in the Tharsis region characterized, we now review documented evidence for the past operation of glaciovolcanic processes in the region.

4. Tharsis Region Glaciovolcanism

In this study, we explore the operation of large-scale surficial lava-ice interactions within the Tharsis region. This assessment is complimented by previous studies which

have proposed and recognized the past activity of other forms of glaciovolcanic activity in the Tharsis region, namely interactions between intrusive volcanism and ground-ice as well as subglacial eruptions. Many past studies have proposed a role for subglacial eruptions and the interaction of intrusive volcanism and ground-ice in the formation of a wide suite of landforms in the Tharsis region including: (1) The circum-Chryse (and other) outflow channels (McKenzie and Nimmo, 1999; Chapman and Tanaka, 2002; Wilson and Head, 2004; Leask et al., 2007a). (2) Chasmata and chaos suggested to be the products of large sub-ice eruptions (Chapman and Tanaka, 2002). (3) Steep-sided plateau-like features with smooth capping layers interpreted to be possible tuyas (Head and Wilson, 2007; Kadish et al., 2008; Martínez-Alonso et al., 2011; Scanlon et al., 2014, 2015). (4) Steep-sided mounds and ridges in the fan-shaped deposits on the flanks of the Tharsis Montes interpreted to be moraine or tundra (Head and Wilson, 2007; Kadish et al., 2008; Scanlon et al., 2014). (5) Steep-sided, leveed flows interpreted to be subglacial lava flows with chilled margins (Head and Wilson, 2007; Scanlon et al., 2014, 2015). (6) Digitate flows interpreted to have interacted with an upslope glacial margin (Head and Wilson, 2007; Scanlon et al., 2014, 2015). (7) Low, widespread mounds suggested to be preserved remnants of effusions of pillow lavas formed during subglacial eruptions (Scanlon et al., 2014, 2015). (8) Sinuous valleys and ridges interpreted to be subglacial eskers and meltwater channels (Head and Wilson, 2007; Kadish et al., 2008; Scanlon et al., 2014, 2015). (9) Various surface materials interpreted to be deposits of tephra and volcanic material derived from subglacial eruptions and the interactions of intrusive volcanic features with ground-ice (Chapman and Tanaka, 2002; Wilson and Head, 2004; Martínez-Alonso et al., 2011).

While this range of features is inferred to have involved a different subset of glaciovolcanic processes from those investigated in this assessment, they support the past occurrence of contemporaneous volcanic and glacial activity and the operation of glaciovolcanic processes in the Tharsis region. Therefore, these features lend additional justification for the glaciovolcanic scenario we explore. With the characteristics of regional volcanism and glaciation within the Tharsis region outlined, and the history of glaciovolcanism now established, the details of the specific glaciovolcanic processes we assess are next outlined.

5. Glaciovolcanic Model Description

Glaciovolcanism is a comprehensive term referring generally to any interaction between volcanic products and water ice (Smellie and Edwards, 2016). In our evaluation of glaciovolcanism in the Tharsis region, we specifically assess a lava-ice interaction geometry involving the emplacement of lava flows atop surficial snow and ice deposits, as proposed in the ice sheet lava heating and loading mechanism (Cassanelli and Head, 2016, 2018a). This particular lava-ice interaction geometry is assessed because, while other lava-ice interaction geometries are possible, supra-ice lava emplacement optimizes heat transfer to the surface ice (Head and Wilson, 2002, 2007; Wilson and Head, 2002a; L. Wilson and Head, 2007; Cassanelli and Head, 2016, 2018a) thereby providing testable maximum end-member estimates for meltwater generation. Furthermore, recent terrestrial studies which have documented the interaction between extrusive lava flows and surface snow and ice on Earth (Edwards et al., 2012, 2013, 2014, 2015) have found the supra-ice geometry to be a dominant mode of lava flow emplacement. Supra-ice emplacement of extrusive lava flows can occur by (1) direct advancement of lava flows onto the ice from

topographically-elevated terrain, as has been observed in glaciated terrestrial volcanic settings ((Edwards et al., 2012, 2013, 2014, 2015), (2) supraglacial eruptions fed by dikes emplaced through surface ice due to high strain rates (Head and Wilson, 2002, 2007; Wilson and Head, 2002a; L. Wilson and Head, 2007) or, (3) eruptions from tuyas that have pierced through surface ice (Smellie, 2007; Kadish et al., 2008; Scanlon et al., 2014).

This particular glaciovolcanic process, supra-ice lava flow emplacement, is predicted to result in significant meltwater production through top-down contact and bottom-up geothermal heating (which could serve as a source for the formation of fluvial features) as well as surface deformation through the subsidence and collapse of lava flows (due to melting and removal of underlying supporting ice) (Cassanelli and Head, 2016, 2018a). We now explore the possible involvement of large-scale, surficial glaciovolcanism through supra-ice emplacement, and the processes predicted to result, in the formation of tectonic and fluvial features in the Tharsis region.

6. Tectonic Features

In our exploratory evaluation of the role of glaciovolcanic processes in the formation of tectonically-related features in the Tharsis region, we focus on the aureole apron formation surrounding the Olympus Mons edifice, and the enigmatic Syria-Thaumasia block/Solis Planum mega-thrust, beginning with the former.

6.1 Volcanic Edifice Aureoles

The Olympus Mons shield edifice (Figure 1) is almost entirely encircled by a scarp of 6-8 km in height (Basilevsky et al., 2005) and the extensive aureole apron units which extend up to ~700 km from the edifice bounding scarp, predominantly off the

northwestern margin (Figure 3a and 3b). The aureole units of Olympus Mons date to ~3.5 Ga (Morris and Tanaka, 1994) and are comprised of at least 9 individual rocky deposits up to 2 km thick characterized by lobate, sub-circular planform and hummocky texture (Tanaka, 1985; De Blasio, 2018) (Figure 3a and 3b). The formation of the aureole units of Olympus Mons has been attributed to a multitude of mechanisms including lava flows (McCauley et al., 1972), subglacial deposits (Hodges and Moore, 1979), ash flows (Morris, 1981), deep-seated deformation (Francis and Wadge, 1983), subaerial landslides (Harrison and Grimm, 2003), subaqueous landslides (De Blasio, 2018), and ice lubricated landslides (Tanaka, 1985). However, consensus among recent analyses of the morphology of the Olympus Mons aureole units suggests an origin by a complex catastrophic landslide (De Blasio, 2018).

There are two plausible ways in which the formation of the aureole units could have resulted from, or been facilitated by, large-scale surficial lava-ice interactions. (1) Subsidence and collapse of an accumulated sequence of Olympus Mons lava flows emplaced atop surrounding ice sheets due to melting (Cassanelli and Head, 2016). (2) Rapid emplacement of lava flows due to basal lubrication from melting of underlying ice, similar in nature to the previously proposed ice-lubricated landslide model (Tanaka, 1985).

Lava flow subsidence/collapse: Formation of the Olympus Mons aureole deposits could plausibly be accounted for by the subsidence and collapse of lava flows emplaced atop ice surrounding the Olympus Mons basal scarp due to the substantial lava flow deformation predicted (Cassanelli and Head, 2016) to result from this process. The deformation of supra-ice emplaced lava flows, resulting from subsidence and collapse

due to melting and removal of underlying ice, would produce a deposit with a heavily disrupted surface morphology potentially similar to that exhibited by the Olympus Mons aureole units (Figure 3).

In order to account for the formation of the Olympus Mons aureole units by the subsidence/collapse of a supraglacial lava sequence, a surrounding ice sheet ~6-8 km thick along the edifice basal scarp is required to produce the elevation offset between the scarp and the surface of the aureoles by melting and removal (Figure 3). The accumulation of an ice sheet to ~6-8 km in thickness would require highly asymmetric, localized ice accumulation and would also likely require an unreasonably large available surface/near-surface water inventory, as this ice thickness exceeds all maximum estimates of martian ice sheet thickness under a wide range of conditions (see Section 3 and Table 2). Moreover, an ice sheet 6-8 km thick, even at the low mean annual surface temperatures at the surface elevation around the base of Olympus Mons (under typical Late Noachian conditions around the 3.5 Ga age of the apron units; (Wordsworth et al., 2013), would be thermally unstable. This can be simply determined by the steady-state one-dimensional heat conduction equation:

$$G = K_T \frac{dT}{dz} \quad (1)$$

where G is the geothermal heat flux (~55 mW/m² during the Late Noachian period; (Solomon et al., 2005), K_T is the thermal conductivity of ice (~2 W/m K under Late Noachian martian conditions; (Cassanelli and Head, 2015), T is temperature (K), and z is depth (m). With a mean annual surface temperature of ~230 K at the base of the Olympus Mons Scarp under typical Late Noachian conditions ((Wordsworth et al., 2013), the

solution of (1) indicates a thermal-equilibrium ice sheet thickness of ~1.5 km. The accumulation of an ice sheet beyond this thermal equilibrium thickness would result in the onset of basal melting. Therefore, unless the rate of ice accumulation greatly exceeds the rate of basal melting (unlikely, as under typical Late Noachian Mars conditions, bottom-up melting rates are comparable to predicted accumulation rates; (Wordsworth et al., 2013; Cassanelli and Head, 2015, 2016), ice sheet thickness could not grow beyond this equilibrium value. If rapid accumulation allowed an ice sheet to grow beyond the equilibrium thermal thickness, the rate of basal melting would consequently increase and act to drive ice sheet thickness down toward the equilibrium value. If the rate of ice accumulation subsequently waned (possibly due to exhaustion of the surface/near-surface ice inventory), the ice sheet thickness would decline back toward the equilibrium value. Therefore, due to the difficulties in the creation and maintenance of an ice sheet of the required 6-8 km thickness, the formation of the Olympus Mons aureole units by the subsidence and collapse of lava flows emplaced atop surrounding surface ice is not favored.

Frictional reduction and lava flow sliding: A second possible surficial glaciovolcanic origin of the Olympus Mons aureole units is by rapidly emplaced supraglacial lava flows (broadly similar to the previously proposed ice-lubricated landslide model; (Tanaka, 1985) in the following manner (Figure 3c): (1) lava flows advance rapidly over surface ice deposits surrounding the Olympus Mons edifice due to formation of a lubricating basal melt lens, (2) once the flows advance beyond the surface ice margin a rampart is produced due to the higher friction basal surface and emplacement deceleration, (3) ice underlying the emplaced lava flow is then melted leading to subsidence and collapse, (4)

surface ice deposits are reformed on the collapsed lava flow, (5) subsequent lava flows are emplaced in a similar manner, back-stepping toward the edifice generating lobate scarps (from the ramparts) and a disrupted surface morphology. This mechanism may also help account for the extensive distances over which the aureole units are deposited away from the Olympus Mons basal scarp (Figure 3a and 3b), as the basal lubrication of supraglacial lava flows by melting has been observed to enhance lava flow emplacement rates under terrestrial conditions (Edwards et al., 2013).

To perform a first-order evaluation of the viability of this mechanism to produce the aureole units, we assess whether a typical Tharsis shield lava flow could reach the 700 km distal end of the aureole units (Figure 3) prior to cooling and solidification. This criteria can be addressed in a very simple fashion by taking a ratio of the timescale required for thermal diffusion across the lava flow thickness (assuming lava flow cooling predominantly by conduction and neglecting advective effects which would enhance cooling due to material movement during emplacement) and viscous flow over the required flow distance (assuming Newtonian rheology for simplicity) which yields:

$$\frac{2 L \kappa \mu}{H^4 \rho g \sin \alpha} \quad (2)$$

where L is a flow distance of interest (~ 700 km in this case), κ is the thermal diffusivity (for which a typical value of 10^{-6} m²/s is adopted), μ is the lava viscosity (for which a value of 10^4 Pa s is used, consistent with measurements from terrestrial basaltic lava flows; (Harris and Allen, 2008), H is the lava flow thickness (for which the typical 50 m value is utilized for Tharsis shield volcanism; Table 2), ρ is the lava density (assuming a typical value of 3,000 kg/m³), g is gravitational acceleration (3.71 m/s² on Mars), and α is

the slope (for which a value of 0.3° is used, the approximate average slope across the longest section of the aureole units). Substituting the relevant parameters into (2) indicates a ratio value on the order of 10^{-5} , suggesting that, for the conditions of interest, viscous flow occurs over a more rapid timescale than diffusive lava flow cooling.

As a second test of the ability of supraglacial lava flows to reach the 700 km distal end of the aureole units, emplacement velocities of terrestrial supraglacial can be utilized. Observations of terrestrial supraglacial flows documents by (Edwards et al., 2012) exhibited an approximate average emplacement velocity of 35 m/hr. To produce an estimate for supraglacial lava flow emplacement velocities on Mars, we scale this emplacement velocity by a factor of $[(\sin \alpha_m / \sin \alpha_e) * (g_m / g_e) * (H_m / H_e)^2]$ (from the Bingham plastic lava rheology) to account for the differences in slope, gravity, and lava flow thickness. Substituting appropriate parameters ($\alpha_m \sim 0.3^\circ$; $\alpha_e \sim 5^\circ$; $g_m = 3.71 \text{ m/s}^2$; $g_e = 9.81 \text{ m/s}^2$; $H_m \sim 50 \text{ m}$; $H_e \sim 5 \text{ m}$; Edwards et al., 2012) yields a scaling factor of ~ 2 which translates to an average emplacement velocity of $\sim 70 \text{ m/hr}$ ($\sim 0.02 \text{ m/s}$). Thus, assuming this as the constant emplacement velocity, $\sim 1 \text{ yr}$ would be required for the lava flow to reach a distance of 700 km. This emplacement timescale is well below the characteristic diffusive cooling timescale for a 50 m thick lava flow (the typical lava flow thickness produced by Tharsis shield volcanism; Table 1), given by H^2 / κ where κ is the lava thermal diffusivity (with a typical value of $\sim 10^{-6} \text{ m}^2/\text{s}$), which is $\sim 80 \text{ yr}$. It is important to note, however, that these simple analyses of lava flow emplacement neglect the influence of advective heat transfer which would result in more rapid lava flow cooling and a consequential reduction in the possible emplacement length. Additionally, the increases in lava viscosity which would occur due to cooling during lava flow

emplacement over this timeframe and scale are not accounted for, a factor which would result in decreases of the emplacement velocity and length. While both of these factors would lead to reductions in the maximum distances over which supraglacial lava flows could be emplaced, it is not uncommon for martian lava flows to exceed 700 km in length, with some individual flows stretching over distances in excess of $1-2 \times 10^3$ km (Keszthelyi et al., 2000, 2004; Warner and Gregg, 2003; Chapman et al., 2010b). Therefore, the supraglacial emplacement of lava flows over the 700 km length of the aureole units is plausible.

We next examine the morphometry of the aureole units as a further test of the glaciovolcanic supra-ice lava flow emplacement origin. Variations in the topography across the surface of the aureole units, produced by the ridges and scarps (Figure 3), is characteristically on the order of ~ 100 m with some scarps showing elevation offsets of up to ~ 1 km (Figure 3). Given the lava flow emplacement origin we assess (Figure 3c), and typical lava flow thickness from Olympus Mons of ~ 50 m, these topographic variations require some form of distal lava flow thickening by a factor of $\sim 2-20$. This could have been accomplished by a rapid change in the resistance to the lava flow from either a change in basal friction as the flow advanced beyond the glacial margins or by encountering a thickened distal rampart from a previously emplaced lava flow (Figure 3c).

Summary: These simple evaluations indicate that, to a first-order, it is possible for the Olympus Mons aureole units to have been formed by the sequential supra-ice emplacement of lava flows. However, given the range of assumptions and simplifications adopted in our analysis, it is worth noting that an alternative possibility to account for the

formation of the aureole units by this process is the recurrent emplacement of lava flows atop surrounding ice sheets without complete melting and removal of the underlying ice. This would require the accumulating lava flows to be relatively thin and the ice to be relatively thick in order for the ice to be preserved through top-down contact heating and melting (Cassanelli and Head, 2016). The supraglacial accumulation of volcanic material could also have been facilitated by episodes of tephra deposition (produced by explosive volcanic activity) contributing additional material and acting to preserve the buried ice from ablation (Wilson and Head, 2009). In this case, a stack of lava flows and tephra with interbedded ice layers could be produced as a result of contemporaneous periods of volcanic material emplacement and ice accumulation. This would produce a scenario exactly analogous to that proposed by (Tanaka, 1985) with a mass of material containing interbedded ice undergoing slow gravity driven spreading due to failure and flow concentrated within the weaker ice layers.

6.2 Syria-Thaumasia Block/Solis Planum Mega-Thrust

The Syria-Thaumasia block of the Tharsis region (Figure 4a and 4b) is a vast, ~3,000 km by ~2,000 km uplifted plateau comprised of volcanic plains. The block is bounded on all margins by zones of fractures and mountains with Valles Marineris to the north, Noctis Labyrinthus to the northwest, Claritas fossae to the west, the Thaumasia highlands wrapping along the east, south, and southeast, and the Coprates Rise to the east (Figure 3a and 3b).

The nature and evolution of the regional tectonics responsible for the formation the Syria Thaumasia block remains enigmatic, and a number of alternative origins have been proposed. (1) Horizontal shortening and compression of the crust and lithosphere, and the

development of an extensive ridge belt due to early volcanotectonic activity and crustal thickening in south Tharsis (Schultz and Tanaka, 1994). (2) Crustal shortening and contractional deformation due to the uplift of the Thaumasia plateau (Dohm and Tanaka, 1999). (3) Gravity-controlled movement/sliding of the Thaumasia plateau facilitated by a ductile lower crustal layer (Webb and Head, 2002). (4) Thick and thin-skinned deformation related to the formation of an orogenic chain with shallow plateau material sliding as an independent block (Anguita et al., 2006). (5) Thin-skinned deformation due to a continental-scale mega-slide on a weak detachment layer comprised of extensive buried deposits of salts and/or mixtures of salts, ices, and crustal debris (Montgomery et al., 2009). (6) Incipient plate tectonism on Mars (Dohm et al., 2013).

A possible surficial glaciovolcanic origin scenario for the Syria-Thaumasia block is thin-skinned crustal deformation along a weak detachment layer of volcanically buried ice (Figure 4d), akin to the mechanisms proposed by (Webb and Head, 2002) and (Montgomery et al., 2009). This glaciovolcanic origin scenario could have resulted from the supraglacial emplacement of the plateau plains-forming lava flows (and potentially explosively generated volcanic tephra; Wilson and Head, 2009)) with incomplete top-down melting of the underlying ice, thereby producing a buried ice layer (Cassanelli and Head, 2016) (Figure 4d). In order for the surface ice to have been preserved through supraglacial lava flow emplacement, the thickness of the plains forming lava flows could not have greatly exceeded ~10 m (Cassanelli and Head, 2016). This is because at these thicknesses the individual lava flows do not contain enough heat energy to melt relatively significant thicknesses of the Noachian/Hesperian regional ice sheets (which are predicted to be ~275 thick on average; Table 2), and as lava flows continue to

accumulate, top-down heat transfer is rapidly attenuated by the presence of intervening layers of previously emplaced chilled lava. The attenuated top-down heat transfer leads to reduced top-down melting during the emplacement of subsequent lava flows, and thus the preservation of the buried ice (Cassanelli and Head, 2016). Typical plains volcanism lava flow thicknesses in the Tharsis region are ~ 20 m (Table 2) and therefore, are relatively consistent with this constraint, allowing the possibility for preservation of surface ice buried beneath supra-glacially emplaced volcanic plains. Additionally, the preservation of surface ice during volcanic burial could have facilitated by the supraglacial deposition of explosively generated volcanic tephra (as was noted in the preceding section) protecting the ice from ablation (Wilson and Head, 2009) and allowing the accumulation of thicker lava flows.

As a result of the limited top-down melting during supra-ice emplacement of ~ 10 m thick lava flows (Cassanelli and Head, 2016), the overall thickness of the Noachian/Hesperian regional ice sheets will not be significantly reduced, and the primary effect of the accumulating lava flows will be the establishment of a thermally insulating layer atop the ice sheets. The additional insulation provided by the accumulating lava flow sequence will act to raise the ice-melting isotherm from depth toward the base of the buried ice, which upon encountering the ice will result in the initiation of basal melting (Cassanelli et al., 2015; Cassanelli and Head, 2016). Depending upon the geothermal heat flux and the rate and total thickness of lava flow accumulation, the onset of basal melting can be deferred beyond the lava flow emplacement period by timescales up to the order of 1 Myr (Cassanelli and Head, 2016). After the onset of basal melting, a meltwater lens would reduce the basal friction and could lead to slip along the basal surface, causing

movement and deformation of the overlying volcanic plains. Deformation of the plains could also occur without the requirement of basal melting by deformation occurring within the weak buried ice layer (Montgomery et al., 2009). In either case, the depth of deformation would be constrained by the thermal equilibrium depth of buried ice in the Tharsis region. This is because after the base of the buried ice surpasses the thermal equilibrium depth due to accumulation of superposed lavas, basal melting of the buried ice would initiate and reduce the ice thickness back toward the equilibrium depth. If superposed lavas accumulated to a thickness sufficient to exceed the ice equilibrium thermal depth, the entire buried ice layer would be subject to bottom-up basal melting by geothermal heat input. It is possible for a thickness of superposed lavas exceeding the ice thermal equilibrium depth to have been accumulated if the rate of construction was more rapid than the rate at which the ice-melting isotherm could ascend toward the base of the buried ice. This “deferred” bottom-up basal melting scenario results from the limitations in total geothermal heat input to the base of the buried ice layer (Cassanelli and Head, 2016). However, given that the maximum timeframe for deferred melting is on the order of 1 Myr, it is unlikely that a thickness of superposed lava greatly exceeding the ice equilibrium thermal depth could have been accumulated.

The ice thermal equilibrium depth can be determined with the use of equation (1) following the procedure outlined in Section 6.1. With a nominal Late Noachian-Early Hesperian mean annual surface temperature in the region of ~ 230 K (Forget et al., 2013; Wordsworth et al., 2013, 2015) and heat flux of ~ 55 mW/m² (Solomon et al., 2005) and an upper-end estimate of the overburden volcanic plains thermal conductivity of 3 W/m K, we find a thermal equilibrium depth of ~ 2.4 km. However, it is possible that the

regional heat flux in the Tharsis region was elevated by volcanic activity (Cassanelli et al., 2015), and that the thermal conductivity of the plains materials was considerably lower. At a regionally elevated geothermal heat flux of 100 mW/m^2 (Cassanelli et al., 2015) and a reduced thermal conductivity value of 2 W/m K , the thermal equilibrium depth is reduced to $\sim 0.9 \text{ km}$. While supraglacial volcanic tephra deposition could have aided in initially preserving surface ice through volcanic burial, the reduced thermal conductivity of the porous tephra material relative to consolidated lava flows would further reduce the thermal equilibrium depths.

Therefore, for the reasons earlier discussed in this section, the depth of deformation in the glaciovolcanic origin scenario for the Thaumasia-Syria block could not greatly exceed these predicted ice thermal equilibrium depths, thus requiring relatively thin-skinned detachment/deformation. These ice thermal equilibrium depths lie at the lower end of the detachment/deformation depth range estimated for the Syria-Thaumasia block deformation which spans from $\sim 1\text{-}15 \text{ km}$ (Anguita et al., 2006; Okubo and Schultz, 2004; Montgomery et al., 2009). The lower end of the estimated $\sim 1\text{-}15 \text{ km}$ detachment/deformation depth range is inferred primarily for the regions wrinkle ridges (Mangold et al., 1998; Montgomery et al., 2009) while the deeper values at the upper-end correspond to the large-scale tectonic features such as Coprates Rise and the Thaumasia Highlands, both interpreted as surface exposures of thrust faults (Plescia and Saunders, 1982; Schultz and Tanaka, 1994; Webb and Head, 2002). Thus, in order for the proposed surficial glaciovolcanic origin (Figure 4d) to account for the complete range of tectonic deformation associated with the Syria-Thaumasia block, and the inferred depths of detachment/deformation, very rapid supraglacial lava (and possibly tephra) accumulation

to a total thickness of ~10 km is required. Given the required accumulation of such a large thickness of volcanic material over a geologically short timescale of ~1 Myr, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by detachment along a volcanically-buried layer of ice (Figure 4d) is not favored.

Summary: We tested a surficial glaciovolcanic origin scenario for the Syria-Thaumasia block by thin-skinned crustal deformation along a weak detachment layer of volcanically-buried ice (Figure 4d). Formation by this mechanism requires accumulation of the volcanic plains which constitute the Syria-Thaumasia block plateau by individual lava flows with thicknesses that do not greatly exceed ~10 m in order to preserve the buried ice (or possibly thicker lava flows if explosive tephra was first deposited onto the ice). This constraint is generally consistent with the estimated lava flow thicknesses for the plains volcanic style responsible for the construction of the plateau (Table 1). Additionally, in order for crustal deformation to have occurred along a weak buried ice layer, the depth of detachment could not greatly exceed the thermal equilibrium depth of ice within the subsurface as, beyond this depth, ice would be unstable to melting and removal. A simple analysis of the thermal equilibrium depth for ice within the Syria-Thaumasia block plateau indicates a maximum depth of ~2.4 km which lies at the lower end of the ~1-15 km depth range estimated for the detachment/deformation of the Syria-Thaumasia block/Solis Planum mega-thrust. However, the lower end of this estimated detachment/deformation depth range is associated with the detachment depths of the regions wrinkle ridge features, while the larger tectonic features such as the Thaumasia highlands and Coprates Rise, both interpreted as surface exposures of thrust faults (Plescia and Saunders, 1982; Schultz and Tanaka, 1994; Webb and Head, 2002), require

deeper detachment toward the higher end of the ~1-15 km range. Therefore, in order for the proposed surficial glaciovolcanic origin (Figure 4d) to account for the complete range of tectonic deformation associated with the Syria-Thaumasia block, and the inferred depths of detachment/deformation, very rapid supraglacial lava accumulation to a total thickness of ~10 km is required. Given the required accumulation of such a large thickness of lava over a geologically short timescale of ~1 Myr, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by detachment along a volcanically-buried layer of ice (Figure 4d) is not favored.

7. Valley Networks

The valley networks (Carr and Clow, 1981; Gulick and Baker, 1989; Baker et al., 1992; Carr, 1996, 2012; Hynek et al., 2010) (Figure 1) are branching systems of linear-to-sinuuous depressions which have persisted as one of the strongest lines of evidence for past fluvial activity on the surface of Mars. Valley networks are widely distributed (Hynek et al., 2010) across the surface of Mars, and have been dated by crater-counting analyses (Fassett and Head, 2008) to predominantly the Late Noachian-Early Hesperian period of Mars history. The valley networks can be divided into two broad populations, a widespread population of systems distributed across the Noachian-aged cratered highland terrains of Mars, and a population of more isolated systems incised into volcanic edifices across the planet. The valley networks associated with the volcanic edifices of Mars typically exhibit younger ages (Fassett and Head, 2008), slightly elevated drainage densities (Hynek et al., 2010), and distinct morphologies (Fassett and Head, 2007) when compared against the older, more ubiquitous valley network systems that occur throughout the cratered Noachian highland terrains. The Tharsis regions contains valley

networks systems from both of these broad populations, with those associated with the broader Noachian highland population located principally within the Thaumasia highlands, and the distinctly younger systems associated with volcanic edifices.

While it is nearly universally agreed that the origin of the valley networks is a result of fluvial activity, whether they formed under relatively warm and wet or cold and icy climate conditions remains unresolved. Accounting for the formation of the valley networks by warm and wet conditions has been problematic due to predictions of climate models suggesting the climate of early Mars was generally cold and icy (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018). These difficulties are amplified for the valley network systems of the Tharsis region because: (1) the Tharsis valley networks systems reside at significant elevations (Figure 1b) in areas predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) to exhibit highly depressed mean annual surface temperatures, and (2) the valley networks incised upon the Tharsis volcanic edifices exhibit relatively young Hesperian/Amazonian ages and were therefore formed in a predominantly cold and dry period of Mars history (Carr and Head, 2010), beyond the timeframe of the potentially warm and wet Noachian. Given these challenges, and the occurrence of the Tharsis valley network systems in a region of intense, prolonged volcanic activity, we examine the possible role of glaciovolcanism in accounting for their formation, an origin which does not require problematic warm and wet climate conditions. We first assess the regional valley network systems of the Tharsis region.

7.1 Regional Valley Network Systems

The regional valley networks of the Tharsis region (Figure 5a and 5b), concentrated within the Thaumasia highlands, are among the most mature fluvial systems on Mars, with systems exhibiting both high drainage densities and stream orders (Hynek et al., 2010). The formation of the regional valley networks in the Thaumasia highlands of the Tharsis region have previously been attributed to: (1) precipitation in a relatively warm and wet climate (Ansan and Mangold, 2006), (2) melting of snow in response to transient heating events in an ambient cold and icy climate (Gulick et al., 1998; Gulick, 2001; Scanlon et al., 2013), (3) groundwater sapping (Mangold and Ansan, 2006), (4) groundwater discharge due to hydrothermal activity (Dohm et al., 1998; Gulick, 1998; Gulick et al., 1998; Tanaka et al., 1998; Dohm and Tanaka, 1999; Gulick, 2001), and (5) groundwater discharge produced by seismic activity from impact events (Gulick et al., 1998; Tanaka et al., 1998; Dohm and Tanaka, 1999). In this assessment we focus specifically on Warrego Valles, a particularly well-developed valley network system (Gulick, 2001) which is incised into the southern face of the Thaumasia highlands (Figure 5a to 5d).

A possible surficial glaciovolcanic mechanism to account for the formation of the Warrego Valles system (and by extension, the other systems of regional valley networks in the Thaumasia highlands), is the dewatering of crust and volcanically-buried ice uplifted to the surface by a Solis Planum mega-thrust (as discussed in the preceding section and originally proposed by (Montgomery et al., 2009) (Figure 5e). Formation of valley networks by this mechanism (Figure 5e) predicts: (1) Valley network heads to begin below the peak elevation of the highland thrust material by a distance equivalent to the depth of the ice melting isotherm below the original surface prior to uplift, as thawed

material would need to be exposed at the surface to cause discharge (this prediction assumes that the uplifted material has not been subject to significant erosion and downwasting). (2) Valley network heads over a narrow range of elevations corresponding to the thickness of thawed subsurface material (which was originally at a depth sufficient to heat the material above the melting point) exposed at the surface by thrusting. (3) Valley networks that exhibit characteristic groundwater sapping channel morphology (as opposed to that characteristic of runoff/overland flow sourced channels) with amphitheater-shaped head walls, and flat floors.

In order to evaluate the consistency of Warrego Valles with the first of these outlined predictions (the initiation of valley heads below the peak elevations of the uplifted host terrain by a distance equal to the depth of the ice-melting isotherm below the original surface) measurements of the distance of the uppermost valley tributary heads from the peak elevations of the host terrain (as indicated by gridded MOLA topography) are taken and compared to estimates of the depth of the ice-melting isotherm below the original surface. The depth of the ice-melting isotherm below the original surface is estimated through the use of the steady-state heat conduction equation following the procedure described in Section 6.1. Assuming a plausible, inclusive range of substrate thermal conductivities from 1-3 W/m K, geothermal heat fluxes from 55-100 mW/m² (Solomon et al., 2005; Cassanelli et al., 2015) and a predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015) mean annual surface temperature in the Thaumasia highlands region of ~230 K yields equilibrium ice-melting isotherm depths of ~0.5-2.4 km below the original surface.

To evaluate if the Warrego Valles system exhibits a relatively consistent band of valley network head elevations, in accordance with the second genetic model prediction, measurements of the elevation at the head of each tributary within the Warrego Valles system are collected using gridded MOLA topographic data (Figure 5f). The histogram of the elevations of the Warrego Valles tributary heads indicates a range of head elevation from ~3-8.5 km, with a majority of the system tributaries (~95 %) heading over a 3.5 km elevation range from ~4-7.5 km (Figure 5f). This compares to the total elevation range of the Warrego Valles system, the elevation difference between the mouth of the main branch to the head of the most elevated tributary (Figure 5c and 5d), of ~2.5-8.5 km. Given the 3.5 km range in valley network tributary head elevations, and the prior criteria requiring a minimum total uplift of ~0.5-2.4 km, in order for valley networks to form by the dewatering mechanism proposed, the crustal materials of the Thaumasia highlands must have been uplifted by a minimum total of ~4-5.9 km. In the area of Warrego Valles, the height of the Thaumasia highlands above the adjacent volcanic plains is ~5 km, but varies from ~1-6 km along portions of the highlands containing valley network systems. Therefore, the total uplift of the Thaumasia highlands is generally within the minimum range required for the valley networks to have formed by the dewatering mechanism, but some localized valley network systems reside in portions of the highlands below the required uplift threshold. Thus, these valley network systems require the operation of a different formation mechanism or substantial erosion, downwasting, or subsidence of the local host highlands terrains.

The final prediction of the dewatering mechanism (Figure 5e) is the production of valley networks exhibiting groundwater sapping morphology, as the source of the water

for the valleys in this scenario is the subsurface. The morphology of the Warrego Valles system, however, is more consistent with that of valleys formed by surface runoff and overland flow (Ansan and Mangold, 2006), suggestive of a precipitation source, either rainfall or snowfall with subsequent melting. Despite the overland flow morphologic characteristics, a rainfall precipitation origin of Warrego Valles has been challenged by the lack of similarly well-formed valley systems in immediately adjacent terrain of the same geologic age (Gulick, 2001), indicating a more localized water source. Therefore, given the lack of sapping channel morphology, and the evidence against a distributed rainfall precipitation source, a favorable alternative mechanism to account for the formation of the Warrego Valles valley network system is the localized melting of surface snow deposits in response to elevated geothermal heat fluxes due to volcanic activity and tectonic uplift (Gulick et al., 1998; Gulick, 2001).

Summary: Using the Warrego Valles valley network system as a case example, we assessed a possible surficial glaciovolcanic origin of the regional highland valley networks concentrated within the Thaumasia highlands of the Tharsis region (Figure 5). Specifically, we evaluated a genetic mechanism for Warrego Valles (and by extension, the other systems of regional valley networks in the Thaumasia highlands) by dewatering of crust and ice uplifted to the surface by a Solis Planum mega-thrust. The formation of the highland valley networks by this mechanism predicts: (1) valley network heads to begin below the peak elevation of the uplifted Thaumasia highlands by a distance equivalent to the depth of the ice melting isotherm below the original surface (as thawed material would need to be exposed at the surface to cause discharge), (2) a narrow range in valley network head elevations corresponding to the thickness of dewatering

subsurface material exposed at the surface by thrusting, and (3) characteristic groundwater sapping channel/valley morphology. The first two of these predicted criteria are met by Warrego Valles, however, the valley morphology of the system is more consistent with an origin by runoff/overland flows associated with an origin by precipitation. Therefore, given the lack of sapping channel morphology, and evidence against a source by rainfall precipitation, a favored origin of the Warrego Valles valley network system is the melting of surface snow deposits in response to elevated geothermal heat fluxes produced by the uplift of heated material from depth during thrusting to form the Thaumasia highlands.

7.2 Volcanic Valley Network Systems

Many of the volcanic edifices within the Tharsis region are incised by systems of valley networks (Figure 6a and 6b). Given that these valley network systems head at or near the summits of the highly elevated Tharsis volcanic edifices (Figure 6a and 6b), it is predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) that, even under even under relatively warm and wet global climate conditions, the formation of these particular valley network systems would have occurred in the presence of local mean annual surface temperatures well below the freezing point of water. Additionally, the volcanic valley network systems of the Tharsis region have been age dated by crater counting techniques (Fassett and Head, 2008; Hynek et al., 2010) to ~0.6-3.7 Ga. These relatively young inferred ages indicate that the formation of these volcanic valley network systems took place beyond the presence of a thicker early atmosphere and the potentially warmer and wetter conditions of the Noachian period, but rather under the cold and dry climate conditions characteristic of the Hesperian and

Amazonian periods of Mars history (Carr and Head, 2010). Therefore, accounting for the formation of the volcanic valley network systems of the Tharsis region by rainfall precipitation is challenging. This difficulty has led to the alternative suggestion for the formation of these volcanic valley network systems by basal melting of snow and ice deposits in response to elevated geothermal heat fluxes near the edifice summits (Fassett and Head, 2006, 2007).

In our evaluation we perform a case study on the Ceraunius Tholus valley network system (Figures 6c to 6f) as it poses the most rigorous test of any formation mechanism. This is because the Ceraunius Tholus valley network system exhibits the highest drainage density ($\sim 0.15 \text{ km/km}^2$) and lowest drainage area ($\sim 1,200 \text{ km}^2$, the upslope area between the heads of each valley) of the volcanic valley networks in the Tharsis region (Hynek et al., 2010) and also exhibits an anomalously young Late Hesperian to Early Amazonian age of $\sim 2.96 \text{ Ga}$ (Fassett and Head, 2008). The valley network system of Ceraunius Tholus (Figures 6c to 6e) is characterized by a radial/parallel drainage structure (clearly structurally controlled by the edifice topography) with low sinuosity valleys reaching a maximum Strahler stream order of 3 and length of $\sim 65 \text{ km}$ (Hynek et al., 2010) and with the degree of fluvial dissection showing an apparent dependence on the flank slopes (Figure 6e).

Prior investigations have suggested the formation of the Ceraunius Tholus valley network systems by basal melting at or near the summit of the edifice in response to elevated peak geothermal heat fluxes of $\sim 220 \text{ mW/m}^2$ produced by an active magma chamber (Fassett and Head, 2006, 2007). At this estimated peak geothermal heat flux, snow and ice deposits of $\sim 400\text{-}600 \text{ m}$ thick would be required to initiate basal melting for

a reasonable surface snow/ice thermal conductivity value of 2 W/m K (Cassanelli and Head, 2015) and predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) local mean annual surface temperatures of ~230 K during the Noachian-Early Hesperian to ~205 K in the Late-Hesperian-Amazonian. (computed by the steady-state heat conduction equation following as described in Section 6.1).

While these thicknesses are well within the range predicted (Fastook et al., 2008) to characterize the tropical mountain glaciers formed in the Tharsis region during the Hesperian and Amazonian periods (Table 2), problems remain regarding the formation of the valley networks by basal melting of surface snow and ice. (1) There is no characteristic morphologic evidence of wet-based/poly-thermal glaciation on the flanks of Ceraunius Tholus that would have likely been produced by basal melting of the requisite ~400-600 m thick ice sheets. Evidence of wet-based glaciation might not have been produced by basal melting of thinner snow packs, but these would likely be too thin to have undergone basal melting. (2) The meltwater production rates produced by bottom-up ice sheet basal melting are generally low (Cassanelli and Head, 2016), making the formation of valley networks by this mechanism alone difficult. The rate of meltwater production per unit area of ice undergoing basal melting can be estimated in a simple manner with the following relationship (Russell and Head, 2007):

$$R = \left[\frac{G}{L_f} - \frac{K \Delta T}{L_f z} \right] \quad (3)$$

where R is the melting rate ($\text{kg/m}^2 \text{ s}$), L_f is the latent heat of fusion of ice ($3.35 \times 10^5 \text{ J/kg}$), z is the thickness of surface snow/ice (m), and the remaining parameters are as previously

defined. Ice sheets accumulating on the Ceraunius Tholus edifice in the presence of an ambient 55 mW/m^2 geothermal heat flux (an average Late Noachian value; (Solomon et al., 2005) and mean annual surface temperatures of 205 and 230 K (representing the Amazonian and Noachian climate conditions; (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018), would reach thermal equilibrium at thicknesses of ~ 2 and 1.5 km , respectively (assuming a constant ice thermal conductivity of 2 W/m K ; (Cassanelli and Head, 2015). If these ice sheets were then exposed to the peak, magmatically-enhanced geothermal heat flux of 220 mW/m^2 at the summit of Ceraunius Tholus (Fassett and Head, 2007), basal melting rates would be $\sim 4.5\text{--}4.8 \times 10^{-7} \text{ kg/m}^2 \text{ s}$ (assuming a constant ice thermal conductivity of 2 W/m K ; (Cassanelli and Head, 2015). Applied over the $\sim 1,200 \text{ km}^2$ summit drainage area of Ceraunius Tholus (the total area above the heads of the individual system valleys), these melting rates yield total melt production rates of $\sim 0.6 \text{ m}^3/\text{s}$. These low total meltwater production rates, generated over an area of $1,200 \text{ km}^2$, are not likely to exceed the infiltration capacity of the substrate or produce substantial erosion of the surface (Cassanelli and Head, 2016, 2018a, 2018b) and are insufficient to supply the discharge needed to form the valley networks of Ceraunius Tholus, the single largest of which is estimated to have involved peak discharges of 4×10^4 to $6 \times 10^5 \text{ m}^3/\text{s}$ (Fassett and Head, 2007). Therefore, the formation of the Ceraunius Tholus valley networks by meltwater production from basal melting of surface snow and ice requires the collection and release of accumulated meltwater reservoirs to produce elevated discharge. This difficulty was previously addressed by invoking the formation and catastrophic drainage of a Ceraunius Tholus caldera lake at elevated discharge rates (Fassett and Head, 2007).

A possible alternative surficial glaciovolcanic formation mechanism for the Ceraunius Tholus valley networks is the direct top-down heating and melting of surface snow and ice by supra-ice emplacement of extrusive volcanic lava flows (Figure 6f) (Cassanelli and Head, 2016, 2018a, 2018c). In this origin scenario, lava flows emplaced atop surface snow and ice can lead to volumetrically substantial meltwater production at rates substantially greater than by bottom-up basal melting (Cassanelli and Head, 2016). Observations of supra-ice lava flow emplacement on Earth (Edwards et al., 2012, 2015) have documented little recognizable post-emplacement evidence to specifically indicate supraglacial lava emplacement. Therefore the lack of any characteristic morphologic signature of this surficial glaciovolcanic process on the flanks of Ceraunius Tholus does not preclude the operation of this mechanism. Assuming a typical Tharsis region individual shield lava flow thickness of ~ 50 m (Table 2), previous models of supra-ice lava emplacement (Cassanelli and Head, 2016, 2018a) suggest peak melting rates on the order of 10^{-4} kg/m² s. Applied over the 1,200 km² drainage area, this melting rate yields a total meltwater production rate on the order of 100 m³/s, several orders of magnitude greater than that for basal melting. While the discharge provided by top-down melting greatly exceeds that of basal melting, reaching the peak discharges required to form the channels incised into the flanks of Ceraunius Tholus (maximally, $\sim 4 \times 10^4$ to 6×10^5 m³/s; (Fassett and Head, 2007) still requires meltwater accumulation and release. A number of possible meltwater confinement and release mechanisms can occur as a result of glaciovolcanic processes leading to sufficiently elevated discharges. These mechanisms, reviewed and assessed in greater detail by (Cassanelli and Head, 2018a, 2018c), include the formation and drainage of surficial supra/englacial meltwater reservoirs. The

formation and drainage of a Ceraunius Tholus caldera lake falls under this broad category of mechanisms, and has previously been demonstrated (Fassett and Head, 2007) to supply sufficient discharges and volumes to account for the formation of the Ceraunius Tholus valley networks.

Summary: The possible surficial glaciovolcanic formation of the volcanic valley networks carved into the flanks of the shield volcanoes within the Tharsis region was evaluated through a case study of the Ceraunius Tholus valley network system (Figure 6). Crater count age dating indicates formation of this system took place during the Late Hesperian to Early Amazonian, ~ 2.96 Ga, a time dominated by cold and dry surface conditions. Due to this relatively young age and the elevations of the host terrain, accounting for the formation of the volcanic valley network systems of the Tharsis region by rainfall precipitation is challenging. As a result of this difficulty, the formation of the Ceraunius Tholus valley network system has previously been attributed to basal melting of surface snow and ice in response to elevated geothermal heat fluxes produced by magmatic activity in the edifice. For a predicted range of elevated geothermal heat fluxes, basal melting of surface snow and ice requires deposits accumulated to thicknesses of ~ 400 - 600 m. However, no characteristic evidence of wet-based glaciation, expected for deposits of this thickness undergoing basal melting, is preserved on Ceraunius Tholus. Moreover, meltwater produced by bottom-up basal melting is generated at a very limited rate and tends to infiltrate into the subsurface. As a result, sustained timescales and special conditions are required to allow meltwater accumulation and release to provide sufficient discharges to erode the Ceraunius Tholus valley network system. As an alternative, we explored an origin of the Ceraunius Tholus valley network system by

surficial glaciovolcanic interactions involving the top-down melting of surface snow and ice by supra-ice lava flow emplacement (Figure 6g). This mechanism is better able to account for the formation of the valley network system by: (1) providing greatly elevated meltwater production rates, and (2) producing no morphologic signature of wet-based glaciation because the surface snow and ice remains cold-based, thereby also preventing major meltwater infiltration. While the surficial glaciovolcanic mechanism is able to provide increased meltwater production rates, accumulation and release is likely still required to supply adequate discharge to erode some of the Ceraunius Tholus system valleys, most plausibly by the formation and drainage of a caldera lake.

8. Outflow Channels

The outflow channels of Mars are enormous erosional channels interpreted to have formed through catastrophic fluvial/diluvial activity (Baker and Milton, 1974; Carr, 1979, 1996; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007). The outflow channels are distinct from the other main group of fluvial features on Mars, the valley networks, in their size, age, and geomorphology. The outflow channels are significantly larger than the valley networks, formed generally later in martian history during the Hesperian and Amazonian periods, originate from discrete point sources as opposed to branching tributary networks, and contain abundant bedform morphology largely absent in the valley network systems (Baker and Milton, 1974; Sharp and Malin, 1975; Masursky et al., 1977; Carr, 1979, 1996, 2012; Carr and Clow, 1981; Baker, 1982; Baker et al., 1992; Clifford, 1993; Carr and Malin, 2000; Wilson et al., 2004; Coleman and Baker, 2009).

Canonical interpretations suggest the formation of the martian outflow channels occurred through catastrophic outbursts of groundwater released from confined aquifers (Baker and Milton, 1974; Carr, 1979, 1996, 2002; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b; Harrison and Grimm, 2004, 2008; Andrews-Hanna and Phillips, 2007) pressurized by excess hydraulic head generated by elevated recharge (Clifford, 1993; Harrison and Grimm, 2004; Russell and Head, 2007; Cassanelli et al., 2015) or the downward growth of a confining cryosphere layer from secular cooling of the martian surface (Carr, 1979, 1996, 2002; Hanna and Phillips, 2005; Wang et al., 2006). Mechanisms proposed to have been responsible for the release of pressurized groundwater to the surface include: intrusive volcanism and dike emplacement (Head et al., 2003b; Leask et al., 2007b), tectonic pressurization and fracturing (Hanna and Phillips, 2006), over-pressurization (beyond the confining lithostatic pressure) by downward cryosphere advancement (Hanna and Phillips, 2005; Wang et al., 2006) or excess hydraulic head (Marra et al., 2014), and the collapse of confining overburden materials (Rodriguez et al., 2005, 2015, 2016).

The Tharsis region is the host of 6 such outflow channels (Figure 1a) (Tanaka et al., 2014) which can be broadly classified into those sourced by fractures/fossae and those sourced by basins/chaotic terrain. Here we specifically examine Mangala Valles (Figure 1a, 7a, and 7b) (Zimbelman et al., 1992; Wilson and Head, 2004; Hanna and Phillips, 2006; Leask et al., 2007a, 2007b; Bargery and Wilson, 2011; Keske et al., 2015), the only clear example of a fracture/fossae-sourced outflow channel in the Tharsis region, and Kasei Valles (Figure 1a, 8a, and 8b) (Baker and Kochel, 1979; Robinson and Tanaka, 1990; Williams et al., 2000; Williams and Malin, 2004; Zealey, 2009; Chapman et al.,

2010a, 2010b; Dundas and Keszthelyi, 2014). Kasei Valles is selected as a case example of the basin/chaos-sourced outflow channels as it is the largest example of this outflow channel class (as well as the largest erosional channel on Mars) and thus offers a challenging test for genetic models. We begin with an assessment of the fracture/fossae-sourced outflow channels through an examination of the Mangala Valles case example (Figure 7a and 7b).

8.1 Fracture/Fossae-Sourced Outflow Channels: Mangala Valles

Mangala Valles (Figure 7a and 7b) is a ~900 km outflow channel system comprised of a series of channels and valleys which emanate from the Mangala Fossa (part of the larger Memnonia Fossae graben system) near the southwest margin of the Tharsis region (Figure 1a). The head of the Mangala Valles system at the Mangala Fossa is ~10 km wide and incised to a depth of ~150 m within the surrounding terrain (Figure 7a and 7b). The outflow channel gradually widens and shallows downstream toward the North where it drains into Amazonis Planitia (Figure 7a and 7b). Paleohydraulic estimates (Leask et al., 2007a, 2007b) indicate that the Mangala Valles outflow channel was formed by a minimum of ~15,000 km³ of water at a peak discharge rate of ~10⁷ m³/s. The geomorphology of the Mangala Valles outflow channel system and the numerous subchannels contained within have been interpreted (Leask et al., 2007a, 2007b) to suggest that, throughout the majority of the duration of the formative flood, the discharge was sustained near the estimated peak values (~10⁷ m³/s), implying a total cumulative formation timescale of ~17 days. Detailed mapping and crater age dating efforts (Keske et al., 2015) suggest that the formation history of Mangala Valles involved at least three

major volcanic events at ~0.7-1, 0.4, and 0.35 Ga, and two major fluvial events at ~0.7-0.8, and 0.35-0.4 Ga, indicating a relatively young Late Amazonian system age.

Due to the characteristics of the Mangala Valles outflow channel system outlined above, accounting for the formation of the system by the canonical groundwater outburst model faces several difficulties: (1) A liquid groundwater system must have been preserved at great depth beneath a very thick cryosphere into the Late Amazonian against efficient vapor diffusive loss to the cryosphere (Clifford, 1991). (2) The groundwater system must have been supplied by a source of active recharge to resupply the aquifer to prevent depletion and to maintain sufficient hydraulic head to fuel discharge (the most plausible source being the Tharsis Rise, still unlikely late in the Amazonian period near the time of formation of the Mangala Valles outflow channel; (Russell and Head, 2007; Cassanelli et al., 2015). (3) The groundwater system must have been saturated to the base of the cryosphere to allow confinement and pressurization, potentially requiring significant volumes of liquid water (Clifford, 1993). (4) Dike emplacement (Wilson and Head, 2002b) or tectonic extension must have produced and maintained an open conduit across the entire ~5-10 km thick cryosphere (Clifford et al., 2010) to allow water outflow to the surface. (5) In order to supply the discharges estimated for the Mangala Valles system the permeability of the aquifer from which the floods originated must have been unreasonably large, far exceeding those common of terrestrial aquifers at similar depths and scales (Head et al., 2003b). (6) Groundwater models predict cumulative erosion of the channel by a series of many smaller individual flooding events (Hanna and Phillips, 2006), inconsistent with the channel geomorphology indicating that the peak flux was sustained throughout most of the duration of channel formation (Leask et al., 2007a,

2007b), implying the formation of the channel by a single, or small number of large floods.

Given these challenges to the traditional groundwater origin model for Mangala Valles, we explore an alternative possible surficial glaciovolcanic formation model (Cassanelli and Head, 2016, 2018a, 2018c) involving supra-ice lava flow emplacement from Mangala Fossa and large-scale lava-ice interactions (Figure 7c and 7d) (similar to that previously applied to the Reull Valles channel in Hesperia Planum and the Athabasca Valles channel in Elysium Planitia; (Cassanelli and Head, 2018a, 2018c). The formation of outflow channels by this glaciovolcanic mechanism (Figure 7c and 7d) is predicted to proceed by meltwater generation and release from both direct top-down contact heating and melting, and by insulation-induced bottom-up heating and melting (Cassanelli and Head, 2016, 2018a, 2018c). Here we disregard bottom-up heating and melting due to the very significant thicknesses of lava and ice required to initiate bottom-up melting, the low rates of meltwater production, and the tendency of bottom-up generated meltwater to infiltrate into the subsurface (Cassanelli et al., 2015; Cassanelli and Head, 2015, 2016).

In order to determine the viability of this glaciovolcanic formation model (Figure 7c and 7d), we estimate the volumes and discharges of water that can be produced by the proposed supra-ice lava flow emplacement mechanism. The volumes and discharges of meltwater that can be produced by this glaciovolcanic mechanism, involving supra-ice lava flow emplacement and top-down contact heating and melting (as proposed in the ice sheet lava heating and loading model; (Cassanelli and Head, 2016, 2018a), depends upon a several factors including: (1) the thickness of surface snow/ice, (2) the thickness of the individual accumulating lava flows, (3) the total thickness of lava accumulated, and (4)

the area over which top-down heating and melting occurs (constrained by the extent of surface ice and lava flows). During the timeframe of outflow formation (Late Amazonian; (Keske et al., 2015), surficial snow and ice deposits in the vicinity of the Mangala Valles channel head are likely to have been related to either the latitude-dependent mantle, or regional ice sheets accumulated at increased planetary obliquity (see Section 3). As a result, the likely range of possible surface ice thicknesses is ~1-100 m (Section 3, Table 2). Volcanism sourced by the Mangala Fossa fracture is interpreted to have involved voluminous, low-viscosity lava flows (Keske et al., 2015), characteristic of flood basalt or plains volcanism. Therefore, the likely thicknesses and extents of individual lava flows emplaced within the Mangala Valles head region are ~20 m and 10^4 km², respectively (Section 2, Table 1). With a typical individual lava flow thickness of ~20 m, complete melting of the thickest surface ice likely to have been present at the Mangala Valles head, ~100 m, is predicted to require the supra-ice emplacement of only two lava flows (Cassanelli and Head, 2016, 2018c). Consequently, the total volume of meltwater which can be supplied by glaciovolcanic interactions of this nature in the Mangala Valles head region is likely to be limited by the amount of surface ice present. Therefore, we assume any surface ice present within the region to have undergone complete melting in response to the emplacement of the volcanic plains, and evaluate the melting area required to supply the volume of water needed to form Mangala Valles over the plausible range of surface ice thicknesses from 1-100 m (Section 3, Table 2). In order to assess the discharge rates which could be produced by supra-ice lava flow emplacement in the Mangala Valles head region, we employ results from previous lava-ice interaction thermal models (Cassanelli and Head, 2016, 2018a, 2018c) for the melting

rates produced during the emplacement of a sequence of two 20 m thick lava flows. Results of prior modeling indicate peak top-down melting rates of $\sim 10^{-3}$ kg/m²s and average rates of $\sim 10^{-4}$ kg/m²s over the course of lava flow emplacement and heat transfer to underlying ice.

We find that, for a range of surface ice thickness from 1-100 m, a total area of melting of $\sim 10^5$ to 10^7 km² is required to supply the minimum 15,000 km³ of water to carve the Mangala Valles outflow channel (Figure 7e) while an area of $\sim 10^7$ to 10^8 km² is required to supply the necessary discharge (10^7 m³/s) (Figure 7f). These required areas of melting are prohibitively large given the $\sim 10^4$ km² extent of the volcanic plains surrounding the head of Mangala Valles. Therefore, in order for the Mangala Valles outflow channel to have been produced by surficial glaciovolcanic interactions and supra-ice lava flow emplacement (Figure 7c and 7d), multiple events of meltwater generation, accumulation, and release are needed to generate the necessary volumes and discharges (as has been found in the assessment of the formation of other outflow channels by this mechanism; (Cassanelli and Head, 2018a, 2018c). A multitude of processes can occur in conjunction with surficial glaciovolcanic interactions and supra-ice lava flow emplacement to result in the confinement and release of meltwater in bursts at greatly elevated discharge rates. These mechanisms and conditions are reviewed and assessed in further detail by (Cassanelli and Head, 2018a) and (Cassanelli and Head, 2018c). The most likely of these processes to have occurred under the conditions relevant to the Mangala Valles system and the surficial lava-ice interaction scenario we investigate, given the regional extent of the surface ice and the supraglacial lava flow emplacement geometry being assessed, are: (1) supra/englacial meltwater accumulation confined by an

ice dam, with release triggered by flotation of the confining ice allowing drainage beneath the dam (Figure 7c and 7d) (much like the processes responsible for the generation of glacial lake outburst floods on Earth, such as the glacial Lake Agassiz events; (Clarke et al., 2004), and (2) supra/englacial meltwater accumulation retained by the ice surface topography, released by subsequent top-down supraglacial overflow (Smellie, 2006) leading to the erosion (by both mechanical and thermal processes; (Marston, 1983; Russell et al., 2001) of a channel in the confining ice (similar to processes observed in some Icelandic jökulhlaup events; (Russell et al., 2001; Smellie, 2006) (Figure 7c and 7d).

Utilizing simple analyses of these meltwater accumulation and release processes described in prior work (Cassanelli and Head, 2018c), we find that the fluxes needed to supply the Mangala Valles could be provided by outflow beneath a confining ice dam through a conduit ~30 km in width, and by a supraglacial sheet flow ~100 km in width, narrowing to ~2 km if a supraglacial channel 50 m in depth were eroded into the ice sheet surface. Therefore, given the ~10 km length of the Mangala Valles channel head, these processes are broadly consistent with the geomorphology of the system (if drainage in the supraglacial scenario was rapidly concentrated into a narrower channel). In order to supply the requisite minimum volume of $15,000 \text{ km}^3$ by surface ice melting over the $\sim 10^4 \text{ km}^2$ area of volcanic plains around the Mangala Valles head, a total of 15-150 melting, accumulation, and release events would need to occur for surface ice thicknesses of 10-100 m. This estimated range of total flooding events required to form Mangala Valles falls within the recurrence range of floods produced by similar processes on Earth (Baker and Bunker, 1985; Waite, 1985; Benito and O'Connor, 2003).

Summary: In order to assess the fossae/fracture-sourced outflow channels of the Tharsis region we performed a case study of the Mangala Valles system, the only clear example of this class of outflow channels in the Tharsis region (Figure 1a and 7). The formation of the Mangala Valles outflow channel has historically been attributed to the catastrophic effusion of groundwater from aquifers confined by an ice-cemented cryosphere. However, due to the relatively young Late Amazonian age of the channel system (~0.35-0.8 Ga), this canonical groundwater outflow channel formation mechanism is problematic, requiring: (1) Preservation of a liquid groundwater system at great depth beneath a very thick cryosphere into the Late Amazonian against efficient vapor diffusive loss to the cryosphere (Clifford, 1991). (2) A source of active recharge to resupply the aquifer to prevent depletion and maintain sufficient hydraulic head for outflow (the most plausible source being the Tharsis Rise, still unlikely late in the Amazonian period near the time of formation of the Mangala Valles outflow channel). (3) Significant volumes of liquid water to saturate the groundwater system to the base of the cryosphere to allow pressurization. (4) Creation and maintenance of an open conduit across the entire ~5-10 km thick cryosphere to allow water outflow to the surface. (5) Unreasonably high aquifer permeability, far exceeding those common of terrestrial aquifers at similar depths and scales. (6) In some models, cumulative channel erosion by a series of many smaller individual flooding events, inconsistent with the channel geomorphology implying the formation of the channel by a single or small number of large floods. Due to these difficulties with the traditional groundwater origin model for Mangala Valles, we explored an alternative possible surficial glaciovolcanic formation model involving supra-ice emplacement of Mangala Fossa lava flows, with subsequent

meltwater accumulation and release by outflow beneath a confining ice dam or supraglacial overflow (Figure 7). We find that, to a first order, this alternative formation model is able to supply the water volumes and discharges needed to erode the channel while conforming to the geomorphic constraints of the system. Therefore, this glaciovolcanic mechanism provides a viable alternative explanation for the formation of the outflow channel system without the difficult conditions required of the classical groundwater model.

8.2 Basin/Chaos-Sourced Outflow Channels: Kasei Valles

Kasei Valles (Figure 8a and 8b) is the largest outflow channel on Mars, situated along the eastern boundary of the Tharsis region (Figure 1a). The channel heads at the Echus Chasma trough of Valles Marineris, extending ~1,600 km to the north where it turns east and continues a further 900 km to drain into Chryse Planitia (Figure 1a). The Kasei Valles outflow channel is estimated to have been eroded by a minimum of $\sim 1.36 \times 10^6 \text{ km}^3$ of water. Prior estimates suggest the channel was formed with peak discharges as high as 10^9 to $10^{10} \text{ m}^3/\text{s}$ (Carr, 1996), while updated estimates (Williams et al., 2000) suggest a lower range of $\sim 8 \times 10^4$ to $2 \times 10^7 \text{ m}^3/\text{s}$ due to improved constraints on the morphometry of interior nested channels.

Kasei Valles is historically interpreted (Robinson and Tanaka, 1990; Carr, 1996) to have been sourced from Echus Chasma through the catastrophic effusion of groundwater or release of an impounded lake. More recent morphologic and mapping investigations (Williams et al., 2000; Chapman et al., 2010a, 2010b), however, indicate that the history of the Kasei Valles outflow channel was far more complex. These studies suggest that Kasei Valles (Williams et al., 2000; Chapman et al., 2010a, 2010b) developed gradually

through multiple flooding events, from multiple sources, over an extended period of time with interspersed periods of volcanic and fluvio-glacial activity occurring from the Noachian into the Amazonian as recently as ~70 Ma. Specifically, unit geomorphology, stratigraphy, and crater age dating suggest a detailed history of the Kasei Valles system as follows (Chapman et al., 2010a, 2010b): (1) Hesperian flood basalt lavas from the Tharsis rise are emplaced at ~3.7 Ga forming ridged plains in the area. (2) At ~3.6 Ga the first event of fluvio-glacial activity occurred with the erosion of the northern, east-trending section of the channel by Early Hesperian-aged floods sourced from the Tharsis Montes area to the west. (3) Several episodes of volcanic and fluvio-glacial activity occurred at ~3.4 Ga, including a separate event of flooding within the northern, east trending-section of Kasei Valles again sourced from Tharsis region to the west, as well as a period of volcanic activity to the south near Echus Chasma and Syria Planum. Contemporaneously, water was being released by numerous channels from the Syria Planum plateau toward Echus Chasma. During this time, the presence of surface ice and ice-rich materials within the Kasei Valles system are suggested by the identification of possible eskers, U-shaped glacial gullies, and glaciovolcanic features such as tinar ridges. (4) Also at ~3.4 Ga, an additional volcanic event resulted in the emplacement of flood lavas within the channel system. (5) From ~1.8-1.0 Ga, glacio-fluvial activity continued within the Kasei Valles system with floods sourced from the Tharsis Rise and Echus Chasma areas developing the north trending portion of the Kasei Valles channel. (6) A prolonged period of volcanic activity in the Late Amazonian lasting until ~89 Ma which involved the emplacement of voluminous lavas sourced from Echus Chasma and the Tharsis Montes. Some of the Amazonian-aged lava flows produced during this period

of activity extend up to 2,100 km in length and exhibit a platy surface morphology. The platy surface morphology bears resemblance to the play-ridged terrain of lava flows in the Central Elysium Planitia region of Mars which have been suggested to be a possible product of surficial glaciovolcanic interactions and therefore an indicator of the presence of surface snow and ice (Cassanelli and Head, 2018c). (7) A final minor phase of fluvio-glacial occurred at ~70 Ma involving small floods from the Echus Chasma trough.

This more detailed understanding of the history of the Kasei Valles outflow channel indicates that the formation of the system involved numerous episodes of volcanic, fluvial, and glacial activity sustained over a large portion of Mars history into the very recent Late Amazonian period. As a result of this history of channel development, accounting for the formation of the Kasei Valles system by the canonical groundwater outburst model faces many of the same issues outlined for the formation of the Mangala Valles system (discussed in Section 8 and 8.1). Given these same difficulties, and the history of Kasei Valles formation (Chapman et al., 2010a, 2010b) indicating sustained coincident volcanic, fluvial, and glacial activity, as well as the possible evidence for glaciovolcanic processes, we explore a possible surficial glaciovolcanic origin of the system. In particular, we evaluate the potential for volcanic activity associated with the construction of the Tharsis region to have interacted with surface ice deposits to generate numerous events of meltwater production and runoff which was collected and erupted from the north-trending portion of Kasei Valles and from Echus Chasma to form the outflow channel (Figure 8c and 8d).

To determine if the Kasei Valles outflow channel system could have been formed by meltwater derived from surficial glaciovolcanic interactions in the Tharsis region (Figure

8c and 8d) we first delineate the total drainage area contributing to the system. The Kasei Valles drainage area was determined using gridded MOLA data at a resolution of ~ 470 m/pixel, yielding a total contributing drainage area of $\sim 6 \times 10^6$ km² (Figure 8e and 8f) (it is worth noting that this drainage area has been derived from modern topography and is likely to have been modified by volcanism, tectonism, and erosion throughout the course of Kasei Valles formation). The drainage basin of the Kasei Valles system delineated through this analysis occupies much of the northeastern Tharsis region (Figure 8e and 8f), areas comprised predominantly of plains and flood basalt volcanic deposits (Figure 2a). Therefore, we assume surficial glaciovolcanic interactions that may have taken place in the Kasei Valles drainage basin to have involved the supra-ice emplacement of lava flows with an average thickness of ~ 20 m (Table 1). Current understanding suggests prolonged development of the Kasei Valles from the Late Noachian to the very recent Amazonian period. Consequently, the Kasei Valles drainage basin may have hosted regional surface ice deposits over a plausible range of thicknesses from ~ 1 -275 m (Table 2).

Given the significant total accumulated thicknesses of lava in this region, on the order of kilometers (Table 1), we assume that any present surface ice within the Kasei Valles drainage basin would have undergone complete melting by surficial glaciovolcanic interactions. Thus, over the total drainage area ($\sim 6 \times 10^6$ km²; Figure 8e and 8f), glaciovolcanic melting of surface ice deposits 1-275 m thick could generate $\sim 5.5 \times 10^3$ - 1.5×10^6 km³ of meltwater. This range of meltwater volumes encompasses the estimated minimum volume of water required to erode the Kasei Valles system, $\sim 1.36 \times 10^6$ km³, which equates to the cumulative melting of ~ 250 m of ice across the Kasei Valles drainage basin. This total cumulative melted thickness of ice is within the range of

thicknesses predicted for surface ice deposits formed within the Tharsis region (Table 2), and could plausibly have occurred through multiple phases of glaciovolcanic melting of relatively thinner surface ice deposits.

Prior analysis (Cassanelli and Head, 2018a, 2018c) of the supra-ice emplacement of ~20 m thick lava flows under relevant martian conditions predicts peak melting rates on the order of 10^{-3} kg/m² and average melting rates over the course of two lava flow emplacement events of 10^{-4} kg/m²s. The emplacement of two 20 m thick lava flows is sufficient to melt the thickest surface ice deposits likely to have been present in the Tharsis region throughout the Amazonian and much of the Hesperian period (~100 m; Table 2). However, in the Late Noachian, regional surface ice sheets in the Tharsis regions are predicted (Fastook and Head, 2015) to exhibit average thicknesses of ~275 m (assuming a total available surface/near-surface water inventory on Mars equivalent to that of the present day; Table 2), a greater thickness than models (Cassanelli and Head, 2018a, 2018c) predict could be melted by sequential emplacement of two 20 m lava flows. As a result, complete melting of surface ice of this thickness would require a greater number of lava flow emplacement events and would therefore decrease the predicted average melting rate. For the level of this analysis, however, we utilize the previously determined average melting rate predicted for the emplacement of two 20 m lava flows noting that the average meltwater production rates produced will represent over-estimates in a Late Noachian scenario of complete top-down surface ice melting by glaciovolcanism.

To derive a maximum estimate of the rate of meltwater production that can be produced by glaciovolcanic melting in the Kasei Valles drainage basin, we assume

simultaneous glaciovolcanic melting across the basin. Under this maximum assumption, application of the predicted peak and average glaciovolcanic melting rates yields total meltwater production rates of $\sim 6 \times 10^5$ - 6×10^6 m³/s. It is, however, improbable that glaciovolcanic melting occurred across the entire Kasei Valles drainage basin during any given time period, with melting likely limited by the extent of individual lava flows emplaced within the catchment area. Given a typical individual lava flow area of $\sim 10^4$ km² in the Kasei Valles drainage basin region (which contains chiefly plains and flood basalt lava flows; Table 1), rates of meltwater production within the catchment are reduced to $\sim 10^3$ - 10^4 m³/s. These values are below those estimated to have been involved in the formation of Kasei Valles, $\sim 8 \times 10^4$ to 2×10^7 m³/s, therefore requiring meltwater collection into reservoirs with subsequent release at elevated discharges. In the Kasei Valles system, meltwater produced within the drainage basin could plausibly have been collected into reservoirs within the southern, smooth portions of the channel (where substantial channel incision is not apparent) by ice-impoundment and within Echus Chasma (Figure 8c and 8d). We define the extent of the possible meltwater reservoir within the north-trending portion of Kasei Valles (Figure 8c and 8d) along the -500 m contour (which very closely follows the edge of the channel) truncated at the juncture between the north-trending and east-trending sections of Kasei Valles where the smooth channel floor morphology transitions to a more highly dissected/incised state (assuming closure of the reservoir at this location by ice-impoundment). The extent of a meltwater reservoir which could be contained within Echus Chasma is defined by the 3.5 km contour which directly traces the boundary of the trough (Figure 8c and 8d). The potential volume of meltwater contained within both of these reservoirs is $\sim 10^5$ km³ each,

thereby requiring a total of ~13 fill and release events to supply the minimum volume of $\sim 1.36 \times 10^6 \text{ km}^3$ needed to erode the Kasei Valles channel. This estimated number of required flooding events is consistent with the recurrence of floods produced by similar processes observed on Earth (Baker and Bunker, 1985; Waite, 1985; Benito and O'Connor, 2003). Under the possible reservoir geometries examined, the $\sim 8 \times 10^4$ to $2 \times 10^7 \text{ m}^3/\text{s}$ discharge inferred (Williams et al., 2000) to have been involved in the formation of Kasei Valles could readily be met given the significant ~400 km width of the possible Kasei Valles channel reservoir (Figure 8c and 8d) (refer to discussion on the effect of reservoir outlet geometry on flood discharge contained in Section 8.1), while the adequacy of floods from Echus Chasma to form Kasei Valles has been established in prior studies (Robinson and Tanaka, 1990; Tanaka and Chapman, 1992).

Summary and implications for other basin-sourced channels: Here we examined the formation of the Kasei Valles system (Figure 8) as an example of the basin-sourced outflow channels in the Tharsis region. We find that the formation of Kasei Valles can be plausibly accounted for by a surficial glaciovolcanic origin involving the generation of meltwater within the Kasei Valles watershed by supra-ice lava flow emplacement with subsequent collection and release from reservoirs formed in Echus Chasma and the southern, north-trending portion of the channel itself (Figure 8c to 8f). This formation mechanism is a natural consequence of the extensive and sustained volcanic and glacial activity in the region and is able to: (1) supply the volumes and discharges of liquid water required for channel formation, (2) account for the distributed sources inferred (Chapman et al., 2010a, 2010b) to have contributed to the system, (3) account for the numerous flooding events suggested by the channel morphology without requiring active

groundwater recharge and a recurring mechanism for cryosphere breaching and groundwater release, and (4) provide a source of abundant liquid water over the prolonged Noachian-Amazonian history of channel development (Chapman et al., 2010a, 2010b) under continuously cold climate conditions (Madeleine et al., 2009, 2014; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016).

In addition to Kasei Valles, a number of other similar basin-sourced outflow channels exist within the Tharsis region, most notably Simud and Tiu Valles (Figure 1a) which appear to emanate from Valles Marineris. The formation of these outflow channels has previously been attributed to the catastrophic drainage of lakes formed within Valles Marineris (Coleman et al., 2007; Harrison and Chapman, 2008; Warner et al., 2013). The surficial glaciovolcanic origin we explored for Kasei Valles offers a viable mechanism to account for the formation of these liquid water reservoirs within Valles Marineris under the cold climate conditions predicted to characterize the majority of martian history (Madeleine et al., 2009, 2014; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). Therefore, the surficial glaciovolcanic outflow channel origin we explore, involving the catastrophic drainage of reservoirs fed by runoff from glaciovolcanic melting of surface ice, offers a compelling explanation for the formation of the basin-sourced Tharsis region outflow channels.

9. Conclusions

The Tharsis region is the largest and most long-lived volcanic province on Mars, with a history of volcanic activity spanning ~4 Gyr. Intensive and widespread volcanic activity began in the region during the Noachian period with the construction of the vast Tharsis rise plateau and continued throughout martian history into the recent Late Amazonian

through further plains volcanism and shield development. Throughout this timeframe of volcanic activity, the Tharsis rise is also predicted to have been a location of regional and local ice accumulation under a wide range of climatic conditions, from the relatively warm and wet conditions often suggested for the Noachian period, to the cold and dry conditions which characterize the current climate. As a result, the Tharsis region of Mars has been a site of coeval volcanic and glacial activity throughout a majority of the geologic history of the planet, raising the possibility for widespread, sustained glaciovolcanic interactions. Here we assessed the possible role of large-scale surficial glaciovolcanic interactions in the formation of several classes of tectonic and fluvial/diluvial features throughout the Tharsis region.

9.1 Tectonic Features

The tectonic features of the Tharsis region assessed in this analysis include the aureole apron units of Olympus Mons (Figure 3) and the Syria-Thaumasia block/Solis Planum mega-thrust (Figure 4).

Edifice aureoles: We find that a surficial glaciovolcanic origin of the Olympus mons aureole apron units (Figure 3) by the sequential emplacement of lava flows (and possible volcanic tephra) emanating from the edifice atop surrounding regional surface ice (Figure 3c) could plausibly account for the morphologic characteristics of the units. Supra-ice emplacement allows rapid advancement of lava flows over the great distances the aureole units are deposited (~700 km), and could result in the construction of the apron ridges through the formation of distal ramparts due to emplacement deceleration by flow beyond surface ice margins or encountering rampart obstacles from prior lava flows (Figure 3c). Gradual construction of the apron units could proceed by the continuation of this process

through successive events of surface ice accumulation and lava flow emplacement. Alternatively, the apron units could have been formed by gravity driven spreading of a mass of interbedded lava flows, volcanic tephra, and ice layers (akin to previously proposed models) formed through a series of supra-ice lava flow emplacement events during which the underlying ice was not completely melting.

Syria-Thaumasia block/Solis Planum mega-thrust: We evaluated a possible surficial glaciovolcanic formation scenario of the Syria-Thaumasia block/Solis Planum mega-thrust (Figure 4) involving detachment and deformation along a weak, volcanically-buried ice layer produced by the accumulation of superposed lava flows (similar to previous models invoking detachment and deformation along a weak buried salt layer) (Figure 4d). Simple thermal analyses indicate that buried ice in the Syria-Thaumasia region would be thermally unstable below ~ 2.4 km and would undergo melting and removal. Therefore, shallow detachment and thin-skinned deformation are required to account for the formation of the Syria-Thaumasia block/Solis Planum mega-thrust by a volcanically-buried ice layer. The shallow required deformation at ~ 2.4 km is generally outside the range of detachment depths estimated to have been responsible for the mega-thrust feature which average to ~ 10 km. Consequently, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by a volcanically-buried ice layer would require rapid accumulation of volcanic materials to thicknesses of ~ 10 km over ~ 1 Myr before bottom-up ice melting could begin. Given the required accumulation of such a large thickness of volcanic material over a geologically short timescale of ~ 1 Myr, formation of the Syria-Thaumasia block/Solis Planum mega-thrust by detachment along a volcanically-buried layer of ice (Figure 4d) is not favored.

9.2 Fluvial/Diluvial Features

The fluvial/diluvial features of the Tharsis region assessed in this analysis include regional highland valley networks, exposed primarily in the Thaumasia highlands, the younger, more isolated volcanic valley networks incised on the regions shield edifices, and the large outflow channels sourced by both fractures/fossae systems and by basins/chaos terrain.

Regional/highland valley networks: Using the Warrego Valles valley network system as a case example, we assessed a possible surficial glaciovolcanic origin of the regional highland valley networks which, in the Tharsis region, are concentrated along the Thaumasia highlands (Figure 5). Specifically, we evaluated a genetic mechanism for Warrego Valles (and by extension, the other systems of regional valley networks in the Thaumasia highlands) by dewatering of crust and ice uplifted to the surface by a Solis Planum mega-thrust (Figure 5e). The formation of the highland valley networks by this mechanism (Figure 5e) predicts: (1) valley network heads to begin below the peak elevation of the uplifted Thaumasia highlands by a distance equivalent to the depth of the ice-melting isotherm below the original surface prior to uplift (as thawed material would need to be exposed at the surface to cause discharge), (2) a narrow range in valley network head elevations corresponding to the thickness of dewatering subsurface material exposed at the surface by thrusting, and (3) characteristic groundwater sapping channel/valley morphology. The first two of these predicted criteria are met by Warrego Valles, however, the valley morphology of the system is more consistent with an origin by runoff/overland flows associated with an origin by precipitation. Given the lack of sapping channel morphology, and evidence against a source by rainfall precipitation, a

avored origin of the Warrego Valles valley network system is the melting of surface snow deposits in response to elevated geothermal heat fluxes produced by the uplift of heated material from depth during thrusting to form the Thaumasia highlands.

Volcanic valley networks: The possible surficial glaciovolcanic formation of the volcanic valley networks, carved into the flanks of the shield volcanoes within the Tharsis region, was evaluated through a case study of the Ceraunius Tholus valley network system (Figure 6). Crater count age dating of this valley network system indicates formation took place during the Late Hesperian to Early Amazonian, around ~2.96 Ga, a time dominated by cold and dry surface conditions. Due to this relatively young age, the formation of the Ceraunius Tholus valley network system has previously been attributed to basal melting of surface snow and ice as a result of elevated geothermal heat fluxes produced by magmatic activity within the Ceraunius Tholus edifice. For a predicted range of elevated geothermal heat fluxes, basal melting of surface snow and ice requires deposits accumulated to thicknesses of ~400-600 m. However, no characteristic evidence of wet-based glaciation, expected for deposits of this thickness undergoing basal melting, is preserved on Ceraunius Tholus. Moreover, meltwater produced by bottom-up basal melting tends to infiltrate into the subsurface and is generated at very limited rates, thereby requiring sustained timescales and melt accumulation to produce sufficient discharge to erode the Ceraunius Tholus valley network system. As an alternative, we explored an origin of the Ceraunius Tholus valley network system by surficial glaciovolcanic interactions involving the top-down melting of surface snow and ice by supra-ice lava flow emplacement (Figure 6g). This mechanism is better able to account for the formation of the valley network system by: (1) providing greatly elevated

meltwater production rates, and (2) producing no morphologic signature of wet-based glaciation because the surface snow and ice remains cold-based, thereby also preventing major meltwater infiltration. While the surficial glaciovolcanic mechanism is able to provide increased meltwater production rates, accumulation and release is likely still required to supply adequate discharge to erode some of the Ceraunius Tholus system valleys, most plausibly by the formation and drainage of a caldera lake.

Fossae/Fracture-sourced outflow channels: In order to assess the fossae/fracture-sourced outflow channels of the Tharsis region we performed a case study of the Mangala Valles system, the only clear example of this class of outflow channels in the Tharsis region (Figure 1a and 7). The formation of the Mangala Valles outflow channel has historically been attributed to the catastrophic effusion of groundwater from aquifers confined by an ice-cemented cryosphere. However, due to the relatively young Late Amazonian age of the channel system (~0.35-0.8 Ga), this canonical groundwater outflow channel formation mechanism is problematic, requiring: (1) Preservation of a liquid groundwater system at great depth beneath a very thick cryosphere into the Late Amazonian against efficient vapor diffusive loss to the cryosphere (Clifford, 1991). (2) A source of active recharge to resupply the aquifer to prevent depletion and maintain sufficient hydraulic head for outflow (the most plausible source being the Tharsis rise, still unlikely late in the Amazonian period near the time of formation of the channel). (3) Significant volumes of liquid water to saturate the groundwater system to the base of the cryosphere to allow pressurization. (4) Creation and maintenance of an open conduit across the entire ~5-10 km thick cryosphere to allow water outflow to the surface. (5) Unreasonably high aquifer permeability, far exceeding those common of terrestrial

aquifers at similar depths and scales. (6) Possibly cumulative channel erosion by a series of many smaller individual flooding events, inconsistent with the channel geomorphology implying the formation of the channel by a single, or small number, of large floods. Due to these difficulties with the traditional groundwater origin model for Mangala Valles, we explored an alternative possible surficial glaciovolcanic formation model (Figure 7c and 7d) involving supra-ice emplacement of Mangala Fossa lava flows, with subsequent meltwater accumulation and release by outflow beneath a confining ice dam or supraglacial overflow. We find that, to a first order, this alternative formation model is able to supply the water volumes and discharges needed to erode the channel while conforming to the geomorphic constraints of the system. Therefore, this glaciovolcanic mechanism provides a viable alternative to explain the formation of the outflow channel system without the difficult conditions required of the classical groundwater model.

Basin/chaos-sourced outflow channels: Our final analysis focused on the basin/chaos-sourced outflow channels of the Tharsis region, utilizing Kasei Valles (Figure 8) as a case example. Kasei Valles is the largest outflow channel on Mars and is interpreted to have developed through numerous episodes of volcanic, fluvial, and glacial activity from multiple sources, over an extended period of time from the Late Noachian to the very recent Late Amazonian. As a result of this history of channel development, accounting for the formation of the Kasei Valles system by the canonical groundwater outburst model faces many of the same issues outlined for the formation of the Mangala Valles system. Given these same difficulties, and the history of Kasei Valles formation indicating sustained coincident volcanic, fluvial, and glacial activity, as well as evidence for glaciovolcanic processes, we explored a possible surficial glaciovolcanic origin of the

system (Figure 8c and 8d). In particular, we evaluated the potential for volcanic activity associated with the construction of the Tharsis region to have interacted with surface ice deposits to generate numerous events of meltwater production and runoff which was collected and released from the north-trending portion of Kasei Valles and from Echus Chasma to form the outflow channel (Figure 8c and 8d). This formation mechanism (Figure 8c and 8d) is a natural consequence of the extensive and sustained volcanic and glacial activity in the region and is able to: (1) supply the volumes and discharges of liquid water required for channel formation, (2) account for the distributed sources inferred to have contributed to the system without requiring multiple sources of groundwater effusion, (3) account for the numerous flooding events suggested by the channel morphology without requiring active groundwater recharge and a recurring mechanism for cryosphere breaching and groundwater release, and (4) provide a source of abundant liquid water over the prolonged Noachian-Amazonian history of channel development under continuously cold climate conditions. Therefore, the surficial glaciovolcanic outflow channel origin we explored (Figure 8c and 8d), involving the catastrophic drainage of reservoirs fed by runoff from glaciovolcanic melting of surface ice, offers a compelling explanation for the formation of the basin-sourced outflow channels of the Tharsis region.

To conclude, we find that glaciovolcanic processes resulting from the extended and overlapping histories of volcanic and glacial activity in the Tharsis region may have contributed to the formation of many of the tectonic and fluvial features in the region offering advantages over previously applied formation models.

Acknowledgements

We gratefully acknowledge support to JWH for participation in the Mars Express High-Resolution Stereo Camera Team (JPL 1237163).

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Volcanic Style	Activity Timeframe	Tot. Area (km ²)	Tot. Thickness (km)	Ind. Area (km ²)	Ind. Thickness (m)
<i>Shield</i>	Noachian-Amazonian	1x10 ⁶	1-10	500 (2-2.5x10 ⁴)	50 (1-220)
<i>Plains</i>	Noachian-Amazonian	2.8x10 ⁸	1-10	10 ⁴ (10 ³ – 2.5x10 ⁵)	20 (1-200)
<i>Flood Basalt</i>	Noachian-Hesperian	4.5x10 ⁶	1-3.5	10 ⁴ (10 ³ – 2.5x10 ⁵)	20 (1-200)

Table 1. Timeframes, areas, and thickness data for the variable styles of volcanic activity which have occurred throughout the history of the Tharsis region. Table entries reflect average/typical values with parenthetical values indicating the range of the associated parameter where relevant. Justification for these data and the associated references are provided in Section 2.

Style	Activity Timeframe	Average Thickness (m)	Thickness Range (m)
<i>Regional</i>	Noachian-Early Hesperian	275-725	0-700 to 0-2,800
<i>Localized</i>	Mid-Hesperian-Amazonian	2,000	100-4,000
<i>Regional</i>	Mid-Hesperian-Amazonian	50	1-100

Table 2. Timeframes and thickness data for the variable styles of glaciation which have occurred in the Tharsis region throughout the history of Mars. Justification for these data and the associated references are provided in Section 3.

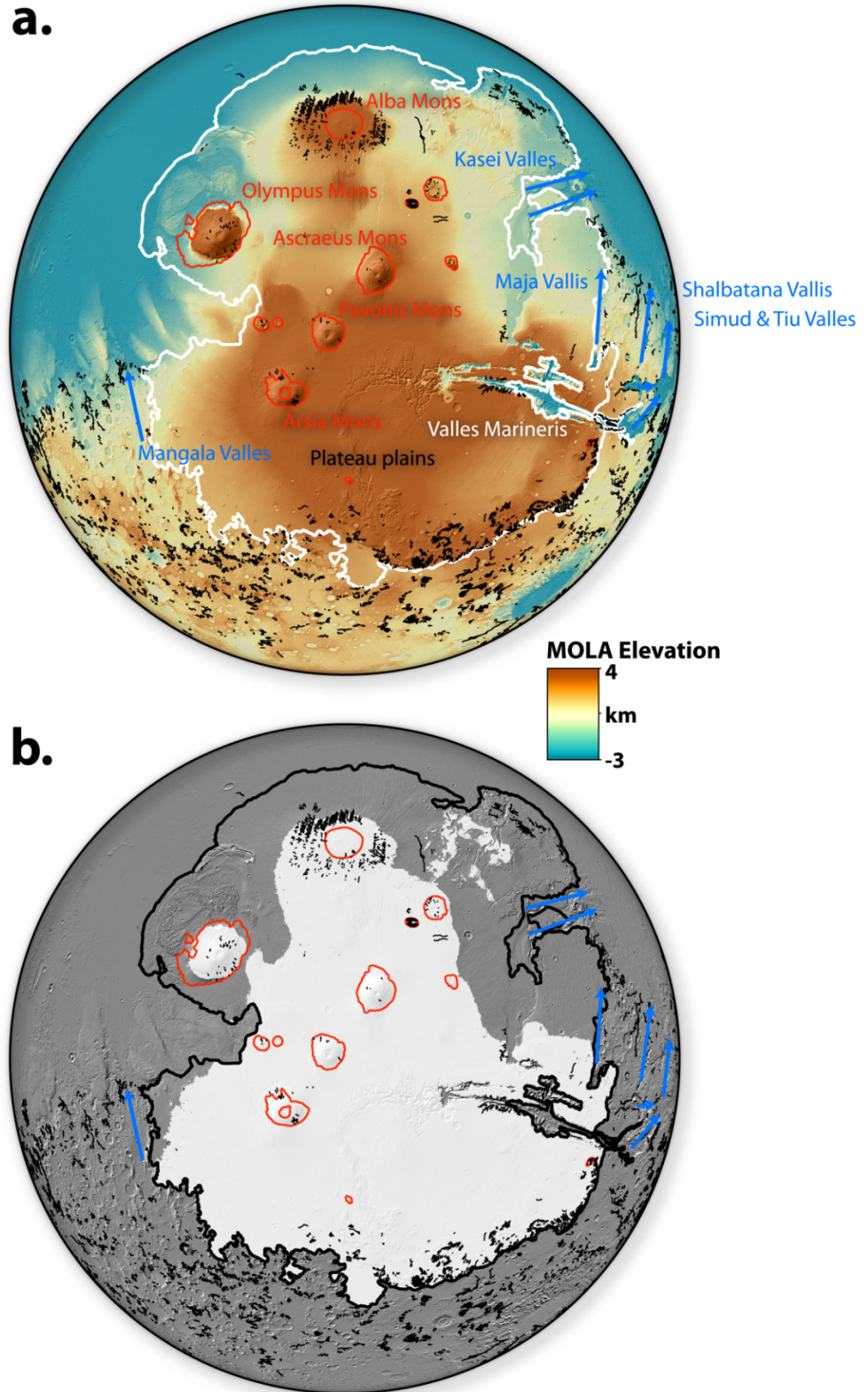


Figure 1. (a.) The Tharsis rise of Mars shown in Mars Orbiter Laser Altimeter (MOLA; (Smith et al., 2001) hillshade data with overlain MOLA topographic data. The defined

extent of the Tharsis region applied in this study (described in Section 2) is shown by the bold white (**a.**) and black (**b.**) lines, major volcanic edifices are outlined and labeled in red, outflow channel flow paths are indicated by blue arrows and labeled, major tectonic features are labeled in white, volcanic plains are labeled in black (Plateau plains including Daedalia, Sinai, Solis, Syria, and Thaumasia Planum), and the valley networks are mapped (Hynek et al., 2010) as black lines. (**b.**) Conceptual representation of the “icy highlands” climate scenario. Areas exceeding a +1 km equilibrium line altitude (in the defined Tharsis region) where snow and ice are predicted (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) to accumulate are shaded white.

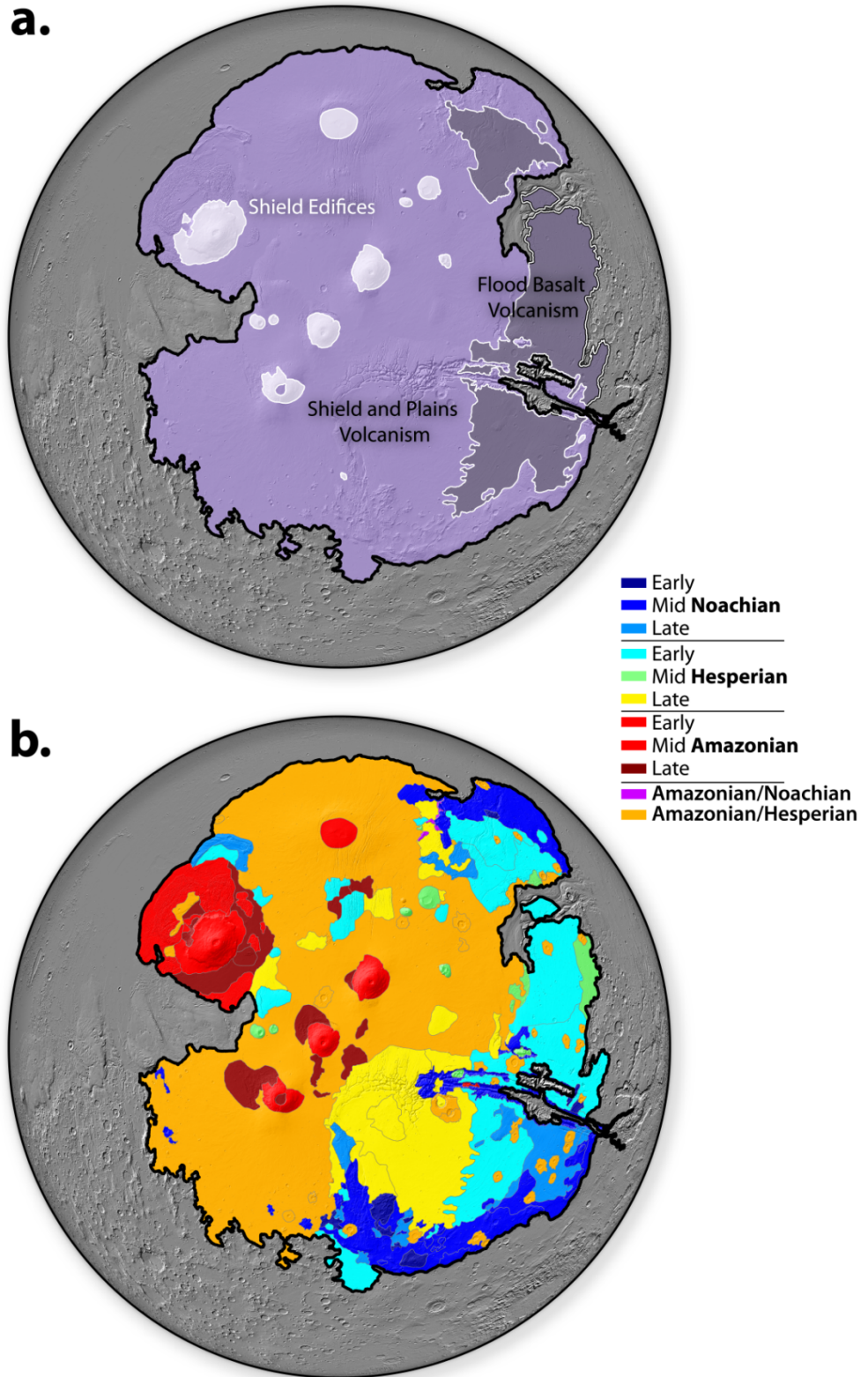


Figure 2. (a.) The distribution of surface volcanic units (Tanaka et al., 2014) generated by the various styles of volcanism which have operated throughout the volcanic history of

the Tharsis region (as defined and discussed in Section 2). **(b.)** Surface ages of mapped (Tanaka et al., 2014) geologic units within the defined Tharsis region. Base data are MOLA hillshade with the extent of the Tharsis region as illustrated in Figure 1 and defined in Section 2.

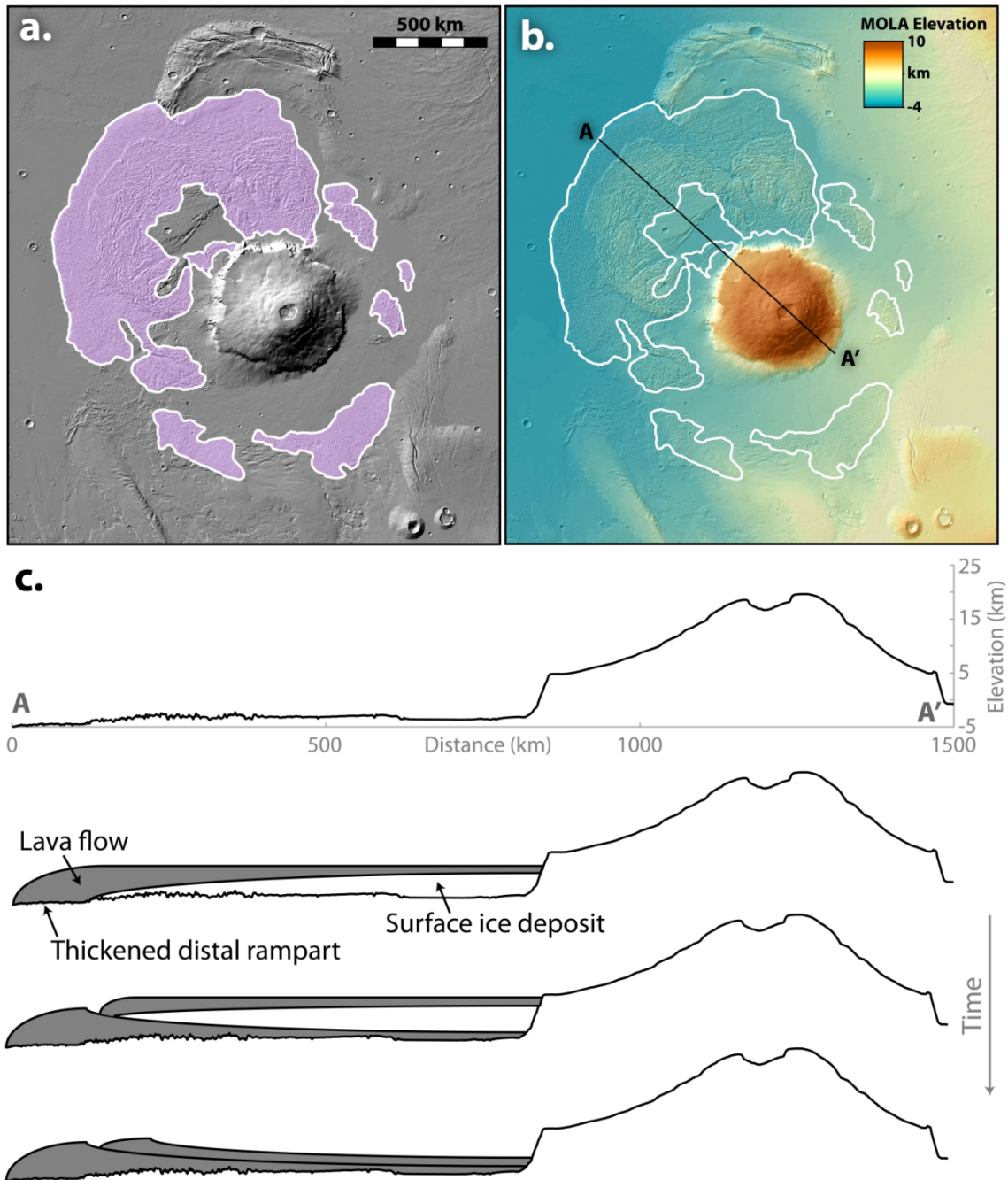


Figure 3. (a.) The Olympus Mons edifice of the Tharsis region with the surrounding aureole formation mapped in the shaded region. (b.) The Olympus Mons construct and surrounding aureole formation shown in shaded MOLA topography. In both panels base data are MOLA hillshade imagery. (c.) A topographic profile across Olympus Mons and the surrounding aureole units with a conceptual representation (not to scale) of possible glaciovolcanic aureole formation mechanism. In the proposed formation scenario, lava flows advance rapidly over surface ice deposits, and produce a rampart when encountering higher friction surface beyond the ice sheet margin. The underlying ice is then melted by the superposed lava flow. Subsequent lava flows emplaced follow a

similar emplacement mechanism, back-stepping toward the edifice and generating the aureole units.

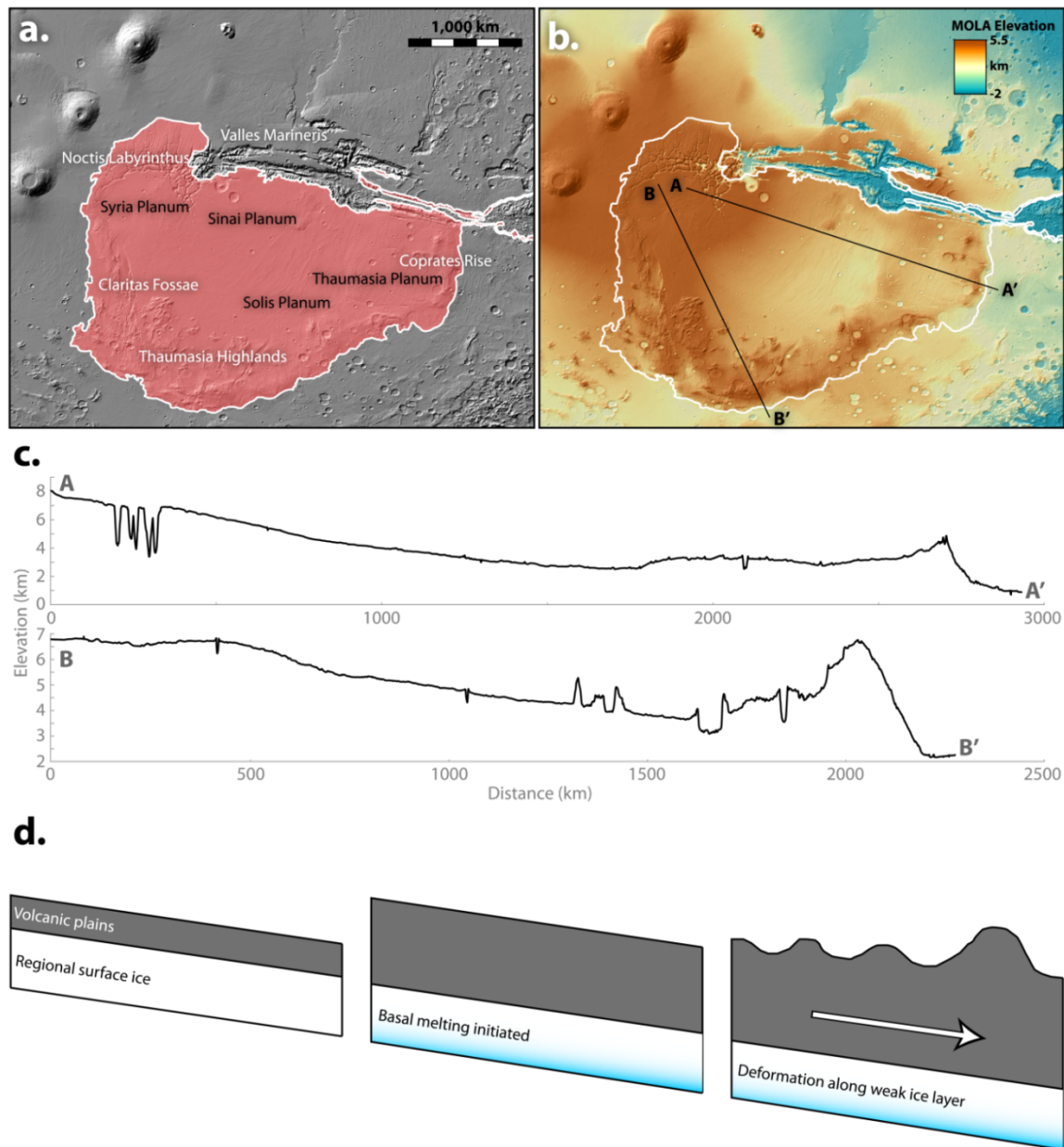


Figure 4. (a.) The approximate extent of the Syria Thaumasia block as defined from the aggregation of mapped (Tanaka et al., 2014) geologic units. (b.) The Syria Thaumasia block and region shown in shaded MOLA topography. In both panels base data are MOLA hillshade imagery. (c.) Topographic profiles across the Syria Thaumasia block. (d.) Conceptual illustration of a possible glaciovolcanic origin mechanism for the Syria-Thaumasia block.

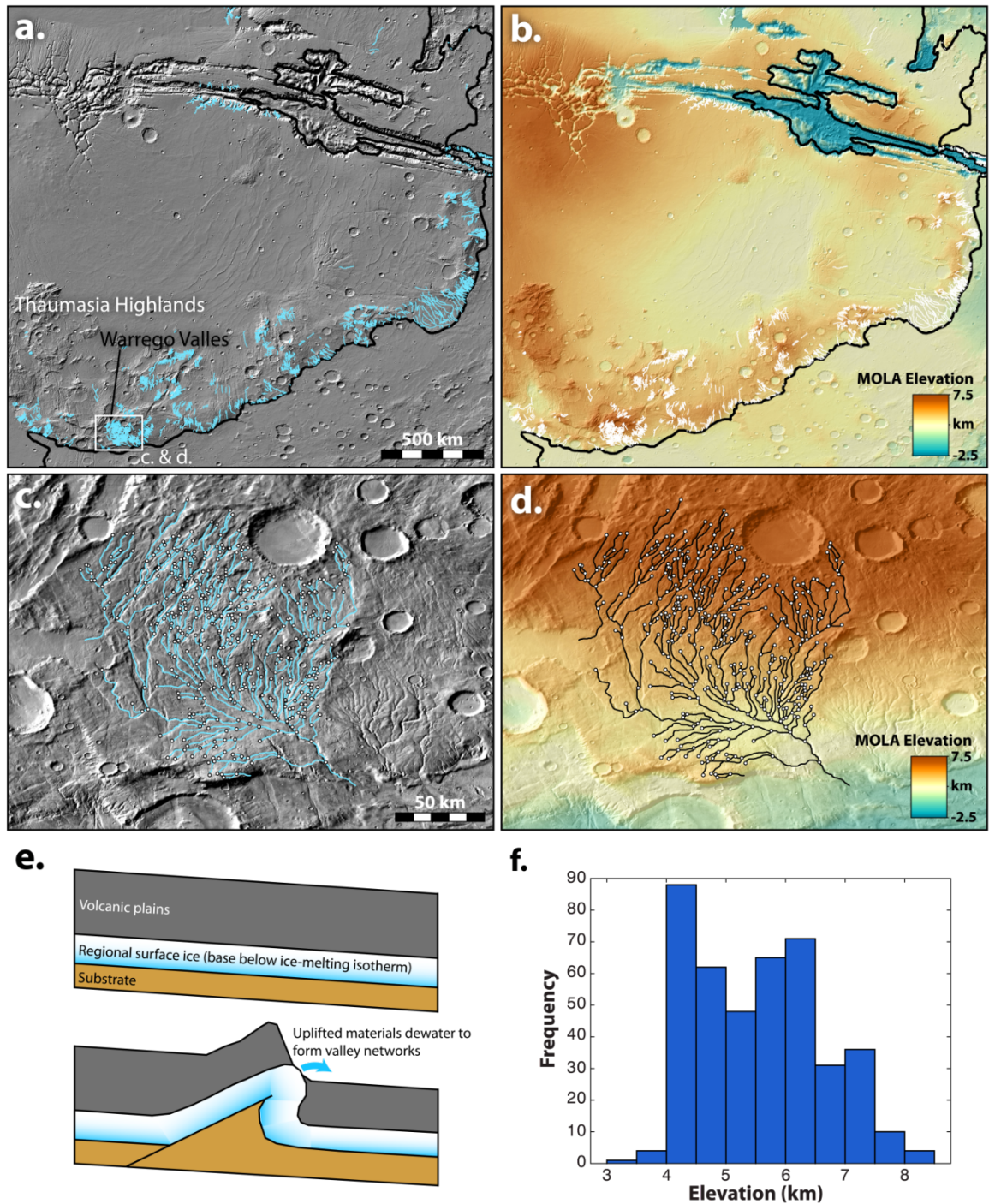


Figure 5. (a. & b.) The regional valley network systems of the Tharsis rise concentrated within the Thaumasia highlands. The valley networks are mapped as the (a.) blue and (b.) white lines. (c. & d.) The Warrego Valles valley network system contained within the Thaumasia highland as indicated by the inset box in panel (a.). Valley networks are mapped as (c.) blue and (d.) black lines with the white dots marking the heads of each individual system tributary. (e.) Conceptual illustration of a possible glaciovolcanic origin mechanism for Warrego Valles and the regional valley networks of the Thaumasia

highlands in the Tharsis region. Formation is proposed to occur by dewatering of geothermally-heated crustal material and buried ice uplifted from an original depth below the ice-melting isotherm. **(f.)** Elevation histogram of tributary heads of the Warrego Valles valley network system marked by the white dots in panels **(c. & d.)**.

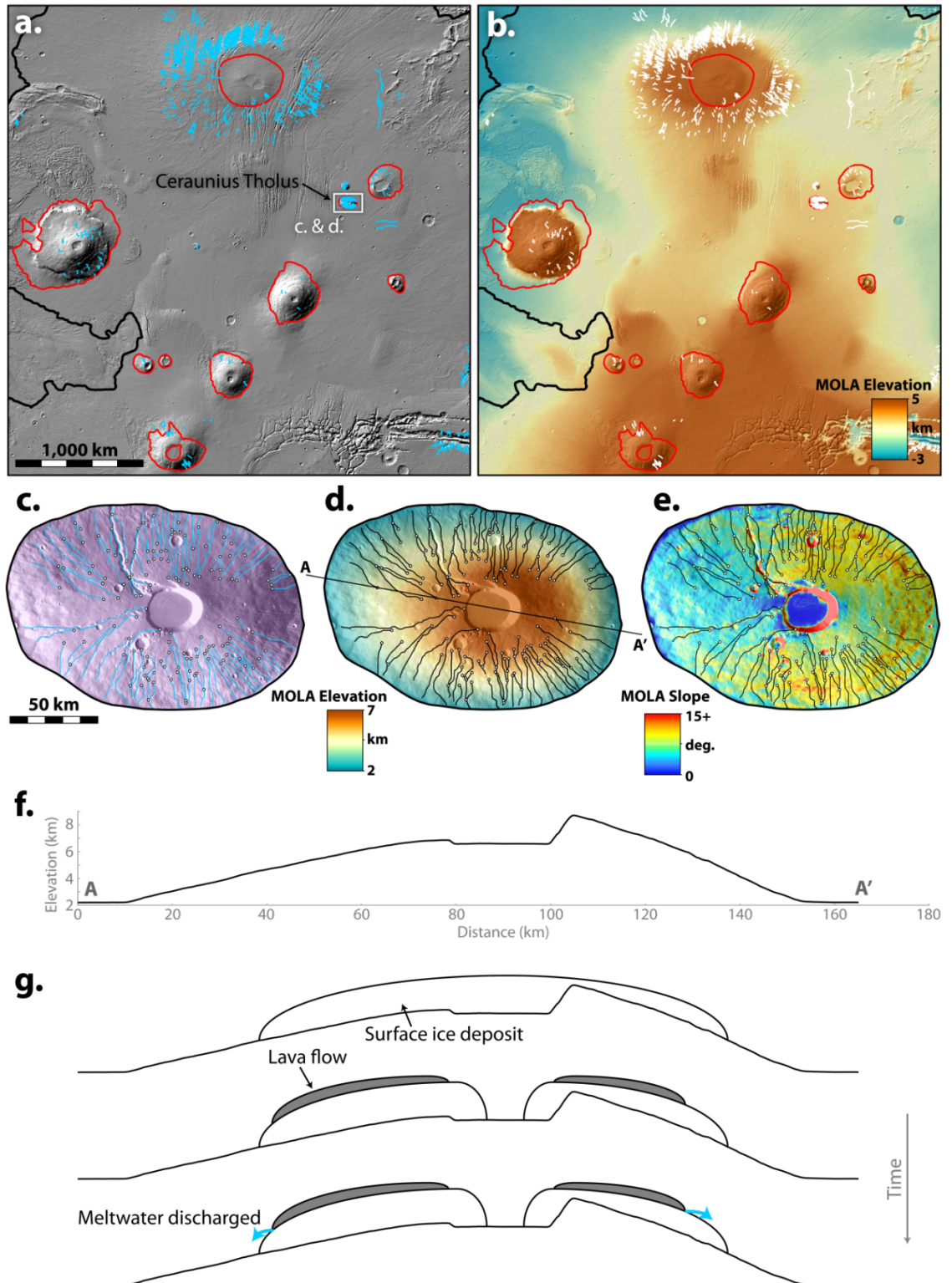


Figure 6. (a. & b.) The volcanic valley network systems of the Tharsis regions concentrated on the flanks of shield edifices. The valley networks are mapped as the (a.) blue and (b.) white lines. (c. & d.) The Ceraunius Tholus valley network system mapped as (c.) blue and (d. & c.) black lines with context provided in panel (a.). (f.) MOLA

Topographic profile, from (d.), across the long axis of the Ceraunius Tholus edifice. (g.)
A possible glaciovolcanic scenario for the origin of the valley networks which are carved into the flanks of the Ceraunius Tholus edifice involving the eruption of extrusive lavas atop surface snow and ice near the summit of the edifice with subsequent melting and discharge.

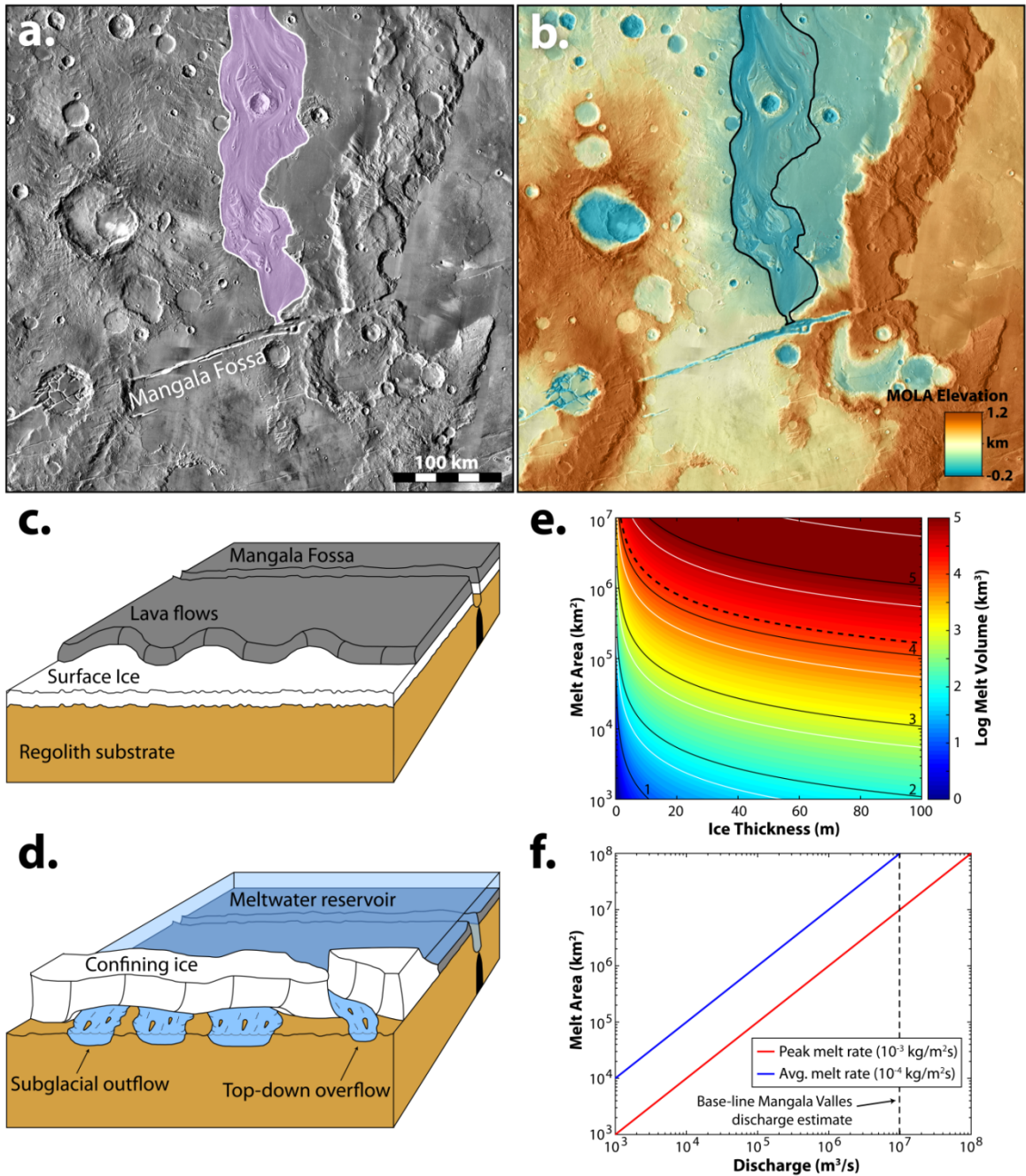


Figure 7. (a. & b.) The Mangala Valles outflow channel system near the southwest margin of the Tharsis region (location context provided in Figure 1a). **(c. & d.)** Proposed glaciovolcanic formation mechanism for the Mangala Valles outflow channel involving **(a.)** the supra-ice emplacement of effusive lava flows generated by the Mangala Fossa leading to **(b.)** meltwater production, supra/englacial accumulation, and subsequent release by outflow beneath a confining ice dam, or top-down overflow. **(e.)** Contours of melt volume produced by lava-ice interactions as a function of the thickness of present regional surface snow and ice and the area of melting. The fine black lines represent integer contours of constant ($\log_{10}$) melt volume with the fine white lines plotted at 0.5

log units between the integer value contours. The nominal range of water volumes ($\sim 15,000 \text{ km}^3$) estimated (Leask et al., 2007a, 2007b) to have been necessary to carve the Mangala Valles outflow channel system are indicated by the bold dashed black lines. **(f.)** The total discharge produced by lava-ice interactions as a function of melting area. The nominal range of discharge estimates (Leask et al., 2007a, 2007b) for the Mangala Valles outflow channel is shown by the dashed black line for reference.

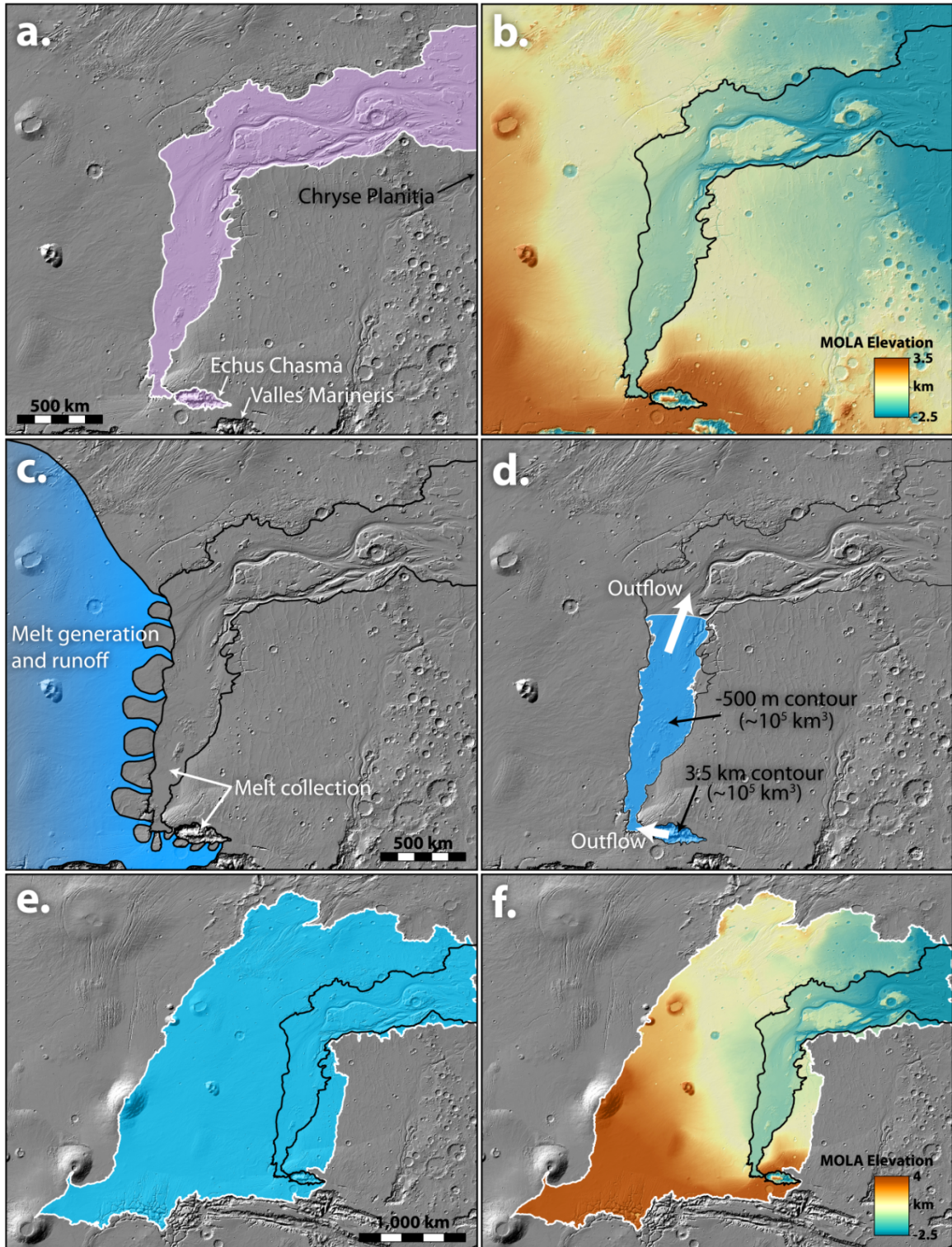


Figure 8. (a. & b.) The Kasei Valles outflow channel along the eastern margin of the Tharsis region (including the Echus Chasma trough of Valles Marineris, a feature closely associated with the Kasei Valles system). **(c. & d.)** A possible glaciovolcanic mechanism to account for the origin of the Kasei Valles outflow channel. Surface snow and ice,

deposited within the elevated Tharsis region, is melted during periods of volcanic activity by glaciovolcanic interactions. Meltwater runoff is then collected within ice-impounded lakes in the north-trending, southern section of Kasei Valles and within Echus Chasma and subsequently release in outflow events to carve the channel system. The extent of meltwater collection within the Kasei Valles channel has been defined along the -500 m contour which closely follows the edge of the channel. The extent of the meltwater accumulation is truncated at the juncture between the north-trending and east-trending sections of Kasei Valles where the smooth channel floor morphology transitions to a more highly dissected/incised state. The extent of meltwater in Echus Chasma is defined by the 3.5 km contour which directly reflects the boundary of the trough. The volume of both of these defined reservoirs is $\sim 10^5 \text{ km}^3$. **(e. & f.)** The contributing drainage basin of the Kasei Valles channel system as determined from present day MOLA topography at a resolution of $\sim 470 \text{ m/pixel}$.

**Chapter 5: Assessing the Formation of Valley Networks on a Cold Early
Mars: Predictions for Erosion Rates and Channel Morphology**

James P. Cassanelli^a & James W. Head^a

^a Department of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

Manuscript under review in: *Icarus*, Manuscript Ref: ICARUS_2018_33_R1

Abstract

The ancient Noachian highlands of Mars host an extensive population of valley networks which formed predominantly during the early geologic history of the planet. Morphologic characteristics of the valley networks have been interpreted to indicate the formation of these features through precipitation-derived fluvial activity and therefore as evidence for a relatively warm and wet Noachian Mars climate. However, these interpretations conflict with the results of sophisticated global climate modeling studies, which suggest that early Mars was dominated by a cold and icy climate with conditions characterized by adiabatic cooling and regional ice sheets within the highlands. Difficulties in replicating the warm and wet early climate conditions interpreted from the valley networks has served as the basis for alternative suggestions for the formation of valley networks by transient heating and snowmelt in an otherwise cold and icy climate. Here we test a conceptual model for valley network formation and incision under cold and icy conditions with a substrate characterized by the presence of an ice-free, desiccated surface regolith and subjacent ice-cemented regolith, similar to that found in the Antarctic McMurdo Dry Valleys on Earth. We implement numerical thermal models, quantitative erosion and transport estimates, and morphometric analyses in order to outline and test predictions for: (1) the nature and structure of the cold and icy Noachian substrate, (2) valley network fluvial erosion and incision rates, and (3) resulting channel/valley morphology. Morphologic predictions are compared against observational data to determine if the valley networks characteristics are consistent with formation in a cold and icy climate. Through this approach we present and develop the underlying conceptual model, identify and generate the fundamental data inputs required to evaluate

it, and perform a preliminary first-order assessment to serve as a basis for further investigation. While these analyses have been performed over a broad parameter space to address relative uncertainties, we report findings with a specific emphasis on the nominal results.

We find that under cold conditions, the substrate is characterized by a kilometers-thick, globally-continuous cryosphere with a ~50-100 meter thick ice-free surface layer in diffusive equilibrium with the atmosphere and underlying ice-cemented regolith, the top of which forms the ice-table. Estimates of the potential infiltration capacity of the ice-free surface regolith layer indicate that surface runoff generation does not require excessively large precipitation rates, and thus does not preclude fluvial activity sourced by rainfall. Additionally, due to the predicted thicknesses of the ice-free surface regolith, any transient period of warming would need to be sustained in duration (~2 kyr) to induce melting of ground ice to initiate solifluction/gelifluction activity. The predicted range of ice-table depths is exceeded by the incised depths of a majority of the valley network population, therefore suggesting interactions were possible. Evaluation of the rates of mechanical substrate erosion and thermal ice-cement erosion by valley network fluvial activity are subject to large uncertainties due to several poorly constrained parameters (*i.e.* the substrate erodibility factor and the valley network fluvial water temperature) but generally indicate that at low water temperatures, consistent with a cold and icy surface environment, thermal erosion of the ice-cement can be outpaced by mechanical erosion of the substrate. In this scenario, the relative efficiency of lateral erosion is predicted to be enhanced, causing preferential channel/valley widening and increased width-to-depth ratios. Assessment of this prediction through a morphometric analysis of valley network

width-to-depth ratios and occurrence of U-shaped cross sections indicates no significant correlation to ice-table depths. This result suggests that either: (1) permafrost was not present in the Noachian Mars substrate, indicating a warmer climate, (2) the permafrost ice-table did not significantly influence valley network width-to-depth ratios, possibly owing to more rapid thermal erosion rates than mechanical, or (3) additional factors not accounted for in this analysis are involved. Potential additional factors which could influence the results of this analysis include variable surface geothermal heat flux, substrate thermal conductivity, substrate composition and physical state as well as valley network preservation state, and morphometric sampling biases (due to data availability constraints). These factors represent areas for future investigation to refine assessments of valley network formation under cold early Mars conditions.

1. Introduction

Mars is host to a diverse array of valley networks (Carr and Clow, 1981; Baker et al., 1992; Carr, 1996, 2000, 2012), systems of linear-to-sinuuous depressions which are widely distributed (Hynek et al., 2010) across the surface of the planet (Figure 1). Valley network systems heavily dissect the relatively older ($>\sim 3.6$ Ga; Werner and Tanaka, 2011) Noachian-aged terrains of Mars (Carr and Clow, 1981; Baker et al., 1992; Carr, 2000; Fassett and Head, 2008; Hynek et al., 2010) but are much more limited in extent within the younger (~ 3 -3.6 Ga; Werner and Tanaka, 2011) Hesperian-aged terrains, restricted primarily to volcanic edifices (Gulick and Baker, 1989, 1990; Gulick et al., 1997; Gulick, 1998; Fassett and Head, 2007). This general stratigraphic distribution, in conjunction with more targeted crater count age dating techniques (Fassett and Head, 2008), indicates that the bulk of valley network formation occurred early in the geologic

history of Mars (~3.7 Ga). The valley networks exhibit characteristics which are indicative of an origin by fluvial activity, providing among the most compelling evidence for the past presence of flowing liquid water at the martian surface. These characteristics include: (1) branching drainage patterns with generally increasing valley size downstream (Carr, 1995, 1996), similar to terrestrial fluvial systems, (2) the presence of alluvial and deltaic deposits within the terminal basins of some valley network systems (Fassett and Head, 2005; Howard et al., 2005), and in some cases (albeit rare) (3) the preservation of bedforms on valley network floors (Gulick, 2001).

While the origin of the valley networks by fluvial activity is almost universally accepted, whether the valley networks formed predominantly by rainfall in a relatively warm and wet early Mars climate, or by transient heating and snowmelt in an ambient cold and icy climate, remains disputed. The formation of valley networks by rainfall in a warm and wet climate has been argued for by geological interpretations and observations which include the widespread presence of degraded Noachian craters in the highlands (which appear to require fluvial erosion) (Craddock et al., 1997; Craddock and Howard, 2002; Forsberg-Taylor et al., 2004; Howard et al., 2005; Irwin et al., 2005b), the need for recharge to account for valley erosion by groundwater sapping processes (Gulick, 2001; Craddock and Howard, 2002), and valley network tributary heads at drainage divides (difficult to explain by groundwater effusion, though in exceptional circumstances, groundwater sapping channels have been observed to breach drainage divides on Earth; Baker, 1982; Gulick and Baker, 1989; Irwin and Howard, 2002). In contrast, the inability of sophisticated climate modeling studies (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) to replicate warm and wet conditions has served as the basis

for suggesting the formation of valley networks by transient heating and snowmelt in an ambient cold and icy climate (Brakenridge et al., 1985; Gulick and Baker, 1989, 1990; Gulick et al., 1997; Gulick, 1998, 2001; Carr and Head, 2003; Halevy and Head, 2014; Head and Marchant, 2014; Palumbo et al., 2018). These climate modeling studies (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) predict that early Mars was dominated by a cold and icy climate characterized by adiabatic cooling and regional ice sheets within the highlands with a mean annual surface temperature of 225K, nearly 50 °C below the melting point of water. Therefore, under these predicted background cold and icy conditions, liquid water would not be stable at the surface, precluding rainfall-derived valley network formation.

Many previous studies have attempted to use the geomorphology of the valley networks to infer the conditions of their formative environment (*e.g.* Gulick, 2001; Williams and Phillips, 2001; Craddock and Howard, 2002; Irwin and Howard, 2002; Hynek and Phillips, 2003; Mangold et al., 2004; Howard et al., 2005; Irwin et al., 2005a, 2005b; Jaumann et al., 2005; Hynek et al., 2010; Hoke et al., 2011; Irwin et al., 2011; Matsubara et al., 2013; Seybold et al., 2018). However, ambiguities in the interpretation of valley network morphology (*e.g.* Craddock and Howard, 2002) have prevented conclusive determination of the prevailing conditions during valley network formation. For example, U-shaped valley cross sections (characterized by steep straight sides and a flat bottom, in contrast to V-shaped valley cross sections), often thought to be indicative of groundwater sapping and thus favoring cold formative conditions, can also be generated by runoff fed valleys in a warm climate (with prime examples contained in

some Hawaiian valley systems) (Gulick and Baker, 1989, 1990; Gulick, 2001; Craddock and Howard, 2002).

Here we revisit this problem by assuming a cold and icy early Mars scenario as predicted by climate modeling studies and investigating the potential influence of these conditions on the formation of the valley networks. Specifically, we test a conceptual model (Head and Cassanelli, 2014) (Figure 2) for channel incision and evolution under cold and icy climate conditions with a substrate characterized by the presence of a dry, ice-free regolith layer and subjacent ice-cemented regolith, the interface between them being defined as the ice-table. This substrate structure is similar to that found in the Antarctic McMurdo Dry Valleys (Head and Marchant, 2014) on Earth, and we later discuss in detail the processes responsible for its establishment as well as its precise characteristics.

In order to quantitatively assess the proposed conceptual model (Figure 2) we evaluate the relative efficiencies of mechanical and thermal fluvial erosion/incision. Mechanical erosion being the removal of substrate material by typical physical fluvial processes (*e.g.* abrasion, cavitation, entrainment) and thermal erosion the melting of pore-filling ice-cement caused by the downward transfer of heat from the fluvial flow into the substrate (a process previously studied in relation to the martian outflow channels by Aguirre-Puente et al., 1994). The conceptual model under assessment (Figure 2) predicts that the characteristic cold and icy substrate structure will cause a discontinuous decrease in vertical mechanical erosional efficiency when the incising channel encounters the more physically resistant, ice-cemented material below the ice-table. The predicted effect of this discontinuous decrease in vertical mechanical erosion efficiency is a

corresponding increase in the relative efficiency of lateral erosion, thus causing preferential channel widening below the ice-table. This prediction depends upon the relative efficiency of the mechanical erosion of the substrate versus the thermal erosion of the ice-cement. If thermal erosion rates are higher than the mechanical erosion rates, the ice-cemented regolith is predicted to melt in advance of the mechanical erosion front and therefore have minimal influence on channel incision and evolution (Figure 2). In contrast, if the thermal erosion rates are lower than the mechanical erosion rates, inhibited vertical erosion is predicted to cause increased relative efficiency of lateral erosion, leading to preferential channel widening (Figure 2). In this latter case, systematically increased channel width-to-depth ratios, and U-shaped channel/valley cross sections are predicted to characterize valley networks which have incised below the ice-table. This behavior, whereby the presence of a permafrost ice-table leads to relatively enhanced lateral erosion, has been well-documented in studies (Costard et al., 2003) of thermal erosion processes in the terrestrial Lena river, a fluvial system residing in the permafrost region of Siberia.

Here we evaluate this conceptual model of valley network formation and the morphologic predictions (Figure 2) (Head and Cassanelli, 2014) through a quantitative assessment of the influence of cold and icy climate conditions, and the associated substrate structure, on fluvial system formation. We implement numerical thermal models and quantitative erosion and transport estimates in order to evaluate: (1) The nature and structure of the cold and icy early Mars substrate; (2) Implications of the substrate structure for runoff generation and fluvial activity; (3) Fluvial incision rates within the predicted substrate above and below the ice-table. Finally, the modeling results are used

to outline predictions for resulting channel/valley morphology. These predictions are then evaluated for consistency with morphologic and morphometric observational data (Williams and Phillips, 2001) to determine if the valley networks exhibit characteristics that are consistent with those predicted for formation under the assumed cold climate scenario. In this contribution we present and develop the underlying conceptual model (Figure 2), identify and generate the fundamental data inputs required to evaluate it, perform a preliminary first-order assessment, and identify sources of uncertainty to serve as a basis for further investigation.

2. Ice-Table Depth

The primary factor predicted to affect valley network channel evolution under cold and icy conditions is the presence of an erosionally resistant ice-cemented regolith within the substrate (Head and Cassanelli, 2014). This substrate structure is established as a consequence of the predicted cold surface conditions on early Mars (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016) which lead to the formation of a kilometers-thick, globally continuous cryosphere (a thickness of substrate continuously below the freezing point of water) (*e.g.* Clifford et al., 2010; Cassanelli and Head, 2015; Cassanelli et al., 2015). In this analysis, we assume the cryosphere to be saturated with pore-filling ice, generating an ice-cemented cryosphere (or permafrost), which could have resulted from freezing of a relict groundwater system (from an earlier warmer period), or from upward vapor diffusion (*e.g.* Clifford, 1991) of groundwater or ice at depth within the subsurface. This assumption of a grain-supported ice-cemented substrate with pore-filling ice not exceeding the available pore space neglects the potential presence of excess ground-ice (ice volume exceeding the available substrate pore space; *e.g.* Bramson et al.,

2015). The potential for excess ground-ice has been disregarded for the purposes of this analysis as this is not the equilibrium condition of the substrate due to vapor diffusive processes acting to reduce excess pore-ice and return the substrate to a grain-supported state (Mellon and Jakosky, 1995; Mellon et al., 1997, 2004; Bramson et al., 2015). Therefore, under these assumed substrate ice-cement conditions, thermal erosion of the pore-ice will not influence physical channel/valley morphometry by widening or deepening. At the surface, a relatively shallow, ice-free, desiccated layer is produced by vapor diffusive equilibrium with the atmosphere (Mellon and Jakosky, 1995; Mellon et al., 1997, 2004), forming the ice-table boundary between the ice-free and ice-cemented subsurface (Figure 2). The depth at which the substrate becomes ice-cemented (the ice-table) is governed by the diffusive equilibrium of water vapor with the atmosphere which depends on mean annual surface temperature, atmospheric water abundance, geothermal heat flux, and the thermophysical properties of the substrate. Given these dependencies, ice-table depths will respond to variations in prevailing climate conditions induced by oscillations in orbital parameters (Laskar et al., 2004) and long-term secular changes in the surface conditions on Mars (due to factors such as atmospheric thinning, increasing solar intensity, and others; Wordsworth, 2016). Here we are interested in evaluating long-term effects of the ice-cement on valley network formation. Therefore, given the relatively short response timescale of ice-table depths (on the order of 1 Myr; Mellon et al., 1997), and the probable lack of significant long-term changes in surface conditions during the timeframe of valley network formation (Fassett and Head, 2008; Hoke et al., 2011), we estimate equilibrium ice-table depths. In considering equilibrium ice-table depths we do not account for potential transient variations due to freezing of infiltrated

surface meltwater which could create a shallow secondary ice-table. These transient ice-tables are not anticipated to greatly affect our results as these features would diffuse away over short timescales (Mellon et al., 1997) relative to that of valley network formation (Hoke et al., 2011). Global equilibrium ice-table depths are calculated with the following relation (Mellon et al., 1997):

$$z_0 = \frac{n_s(T_0) - n_{atm}}{\left. \frac{\partial n_s}{\partial T} \right|_{T_0} \frac{\partial T}{\partial z}} \quad (1)$$

where z (m) is depth below the surface, z_0 is the equilibrium depth of the ice-table, n_s (molecules/m³) is the saturation vapor density of water, n_{atm} is the mean atmospheric near-surface vapor density (molecules/m³), T is temperature (K), and T_0 is the temperature at depth z_0 . Solution of equation (1) for the equilibrium ice-table depth at any surface point requires the mean annual surface temperature, mean atmospheric water abundance, and geothermal gradient ($\frac{\partial T}{\partial z}$, which depends on substrate thermal conductivity and geothermal heat flux) at that point. To perform this calculation we utilize mean annual surface temperature (Figure 3a) and atmospheric water abundance data (Figure 3b) (translated into mean atmospheric near-surface vapor density following Mellon et al., 2004) from recent global climate model studies (Palumbo et al., 2018) under typical predicted Late Noachian climate conditions (1 bar CO₂ atmosphere and 45° obliquity; Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016). We test a range of geothermal gradients by assuming a global average Late Noachian geothermal heat flux of 55 mW/m² (Solomon et al., 2005) and assessing a range of regolith thermal conductivity values of 0.1-2 W/m K (Mellon et al., 1997) to account for

variability and uncertainty in the regolith properties. The saturation vapor density of water, n_s , is calculated from an empirical relationship following Mellon and Jakosky (1993). The major parameter values utilized to perform this and following computations, as well as the parameter space explored, are summarized in Table 1.

We find that over the range of tested substrate thermal conductivity values, equilibrium ice-table depths in the valley network regions (Figure 4) range from several meters to greater than 400 m. Nominal results are represented by a porous regolith substrate with a thermal conductivity of 0.5 W/m K. From a thermophysical standpoint, a porous regolith material is comparable to a porous sedimentary material (with thermal conductivity depending primarily on porosity) and thus the adopted nominal thermal conductivity value is additionally representative of sedimentary rocks. The abundant presence of these particular surface materials across Mars (*e.g.* Malin and Edgett, 2000; Edwards et al., 2009) supports the adoption of this nominal substrate parameter. Lower and upper bound estimates are given by fine particulate and dense rock substrates with corresponding thermal conductivity values of 0.1 and 2 W/m K, respectively. Results for the nominal substrate indicate a global range of equilibrium ice-table depths from ~50-100 m (Figure 4) with the global distribution (Figure 5a) showing a general trend of decreasing ice-table depth with altitude in response to the decreasing temperatures caused by adiabatic cooling. In the valley network regions (Figure 4 and 5b) equilibrium ice-table depths exhibit the same total range of ~50-100 m and average to ~80 m.

The ice-table depths predicted by this analysis have several important implications. (1) The range of incised depths of many valley network systems (Figure 4) (Williams and Phillips, 2001) exceed the range of predicted ice-table depths, and therefore valley

network channels may have interacted with the substrate ice-table. (2) Ice-table depths on the order of tens of meters suggest that any meltwater generated at the surface through transient heating could infiltrate (and freeze) and mechanically erode the substrate for an extended period before reaching the ice-table. (3) Any warm periods would have to be of extended duration (~ 2 kyr) in order for the surface temperatures to penetrate the ~ 50 - 100 m depth to the ice-table to cause extensive melting of ground ice and induce solifluction or gelifluction activity. (4) Given the relatively large predicted ice-table depths (Figure 4 and 5), the desiccated, ice-free substrate layer could efficiently absorb precipitation caused by any punctuated surface heating events. Consequently, very large precipitation rates could be required to generate runoff and fluvial erosion, suggesting fluvial formation by processes other than precipitation. A viable alternative process being the melting of surficial snow and ice deposits (Gulick, 1998, 2001) that could allow melt to collect along the snow/ice surface during the course of drainage (due to the relative impermeability of surficial snow/ice deposits; Cassanelli and Head, 2015) producing elevated discharge/runoff rates that could exceed the substrate potential infiltration rate (after melt drainage beyond the snow/ice margins). This, in turn, would provide support for a cold early Mars climate scenario. In order to assess this important final implication, we now evaluate the potential infiltration capacity of the martian regolith in order to determine the rates of precipitation required to generate runoff.

3. Potential Infiltration Rates

The potential infiltration rate (defined as the maximum rate at which water can be accepted into a porous material) of regolith ultimately depends upon the physical and hydraulic properties of the material. However, the properties of the martian regolith are

poorly characterized due to the lack of available in situ measurements. This is compounded by our interest in evaluating the infiltration capacity of the Noachian martian regolith, which has since undergone several Gyr of modification. Therefore, in the absence of reliable in situ regolith data, we attempt to identify a representative terrestrial soil through a grain size distribution analysis of a range of martian regolith simulants (Figure 6) (Skaggs et al., 2001; Seiferlin et al., 2008; Zeng et al., 2015) (though potential exists for future improvement of these estimates through the incorporation of additional regolith grain size distribution constraints including in-situ measurements and calibrated remote sensing mapping using thermal inertia and albedo; *e.g.* Williams et al., 2013; Jones et al., 2014; Harrison et al., 2015). The grain size distributions of the Mars regolith simulants are relatively consistent, with the bulk of the material comprised of silt and sand-sized fractions with a smaller clay-sized component (Figure 6). This grain size distribution is approximately equivalent to that of a terrestrial clay loam. We therefore adopt a clay loam soil as a nominal representative physical analogue to the Mars regolith. To address the large uncertainties involved in this analogue analysis, we also explore a broader range of soils dominated by finer and coarser fractions from clay to coarse sand.

Hydraulic data for the representative clay loam analog and the range of explored terrestrial soils (Chow et al., 1988) are utilized to estimate infiltration rates using the Green-Ampt method according to the following relations:

$$F(t) = K_h t + \psi \Delta\theta \ln \left(1 + \frac{F(t)}{\psi \Delta\theta} \right) \quad (2)$$

$$f(t) = K_h \left(\frac{\psi \Delta\theta}{F(t)} + 1 \right) \quad (3)$$

where F is the cumulative infiltration depth (m), t is time (s), K_h is the saturated hydraulic conductivity (m/s), ψ is the capillary head (m), $\Delta\theta$ is the unsaturated porosity, and f is the infiltration rate (m/s). The hydraulic data for the terrestrial soil analogs used to estimate infiltrations rates includes the capillary head (ψ), unsaturated porosity ($\Delta\theta$), and saturated hydraulic conductivity (K_h). For each of these parameters, an average value was computed from typical parameter ranges (using data reported in Chow et al., 1988) for each of the considered soil types (from a fine-dominated clay, to a coarse sand, including the adopted representative clay loam analog). The lower gravitational acceleration on Mars is accounted for by scaling the saturated hydraulic conductivities of the soils which are directly proportional to gravity in accordance with the following relation:

$$K_h = \frac{k \rho g}{\mu} \quad (4)$$

where k is the soil permeability (m^2), ρ is the density of water (kg/m^3), g is gravity (m/s^2), and μ is the dynamic viscosity of water (Pa s). Hydraulic conductivity values have been adopted assuming fresh water properties (*i.e.* density and viscosity). While this assumption neglects the potential effects of high dissolved solid content (*e.g.* brine), it is improbable for infiltrating water produced through either precipitation or snow/ice melting to have incorporated a large dissolved load before undergoing infiltration. Therefore, the assumed water properties are not likely to have any significant influence on the estimated hydraulic conductivity values. The parameter values utilized, and ranges considered (through consideration of the soil types of varying coarseness) for each of these hydraulic properties are indicated in Table 1.

Infiltration rates are computed by solving for the cumulative infiltration depth with equation (2) through successive substitution using an initial value of $F(t)$ set to $K_h t$. After convergence is reached, the final value of $F(t)$ is used to compute the infiltration rate $f(t)$ by equation (3). Infiltration rates have been calculated over a plausible total precipitation event timescale of 12 hours (Figure 7), though the precise duration is not of significant importance as infiltration rates quickly decay toward steady-state values. The potential infiltration rates begin at initially elevated values (Figure 7) because the soil is assumed to be dry (or to contain only an irreducible/residual water content) prior to precipitation. Thus at the onset of precipitation, capillary action exerts a force drawing water into the material matrix and increasing the infiltration rate. As infiltration proceeds, the material becomes increasingly moistened, reducing the force of capillary head. With sufficient time the capillary contribution becomes negligible and the potential infiltration rates reach steady-state values equal to the saturated hydraulic conductivity of the soil being evaluated.

The Green-Ampt method we employ to calculate infiltration rates is subject to a number of limiting assumptions (Talbot and Ogden, 2008; Ogden et al., 2015). (1) The method, as implemented here, assumes that there is no ponded surface water which would impart an additional hydraulic head and increase infiltration rates. However, significant ponded water depths are not likely to have been achieved in the sloping topography of the valley network regions of interest (Figure 1 and 5). In these regions surface runoff is likely to have dominated, thus minimizing the limitations of this assumption. (2) The Green-Ampt method inherently assumes a sharp wetting front boundary and, as a result, assumes uniform soil hydraulic properties and soil moisture

content. Therefore, any variability in hydraulic properties with depth that might influence infiltration rates are not accounted for. However, given the inherent uncertainties in the martian regolith properties, this assumption does not contribute any additional limitations to our analysis.

We find that potential infiltration rates for the representative terrestrial clay loam quickly decay to ~ 1 -10 mm/hr over the plausible 12 hr precipitation event timescale investigated (Figure 7). At these potential infiltration rates, runoff could be generated at light to heavy terrestrial rainfall rates (aided by the reduction in the hydraulic conductivity due to the lower martian gravity). Therefore, if the adopted terrestrial clay loam analogue soil is representative of the Noachian martian regolith, runoff generation on Mars would not require prohibitively large rainfall rates. This is consistent with morphometric evaluations (Seybold et al., 2018) of valley network bifurcation angles indicating a predominance of overland flow (without a strong contribution from groundwater flow) though does not distinguish a warm/wet or cold/icy formative environment.

Considering an expanded range of soil examples indicates infiltration rates of ~ 1 mm/hr for a clay dominated soil, and ~ 100 mm/hr for a sand dominated soil (Figure 7). Thus, if the Noachian martian regolith contained a more significant clay component, runoff could be generated readily by light rainfall rates, while a sand dominated regolith would require violent rainfall/deluge conditions. We conclude that for a nominal regolith material, runoff generation on Mars does not require unreasonably large precipitation rates and return to our analysis of valley network formation, first examining mechanical erosion rates.

4. Mechanical Erosion Rates

The rate of channel incision (Figure 2) during valley network formation is governed by the mechanical erosion rate produced by fluvial flow. In order to estimate the mechanical fluvial incision rates achieved during valley network formation, we implement a simple stream power based estimate (Sklar and Dietrich, 1998) following Hurwitz et al. (2010) and Hurwitz et al. (2012), with the vertical erosion rate ($\frac{\partial z}{\partial t}$ in m/s) given by:

$$\frac{\partial z}{\partial t} = K_s \rho g Q \sin(\alpha) \quad (5)$$

where K_s (Pa^{-1}) is a substrate erodibility factor expressing the efficiency of the incision process, Q is the average volume flux per unit channel width (m^2/s), α is the channel floor slope ($^\circ$), and all other parameters are as previously defined. Estimates of the vertical mechanical erosion rate produced by equation (5) are sensitive to the value of the erodibility factor K_s (Hurwitz et al., 2010), which is a poorly constrained parameter aggregating the effects of many factors (Sklar and Dietrich, 2001) (*e.g.* substrate strength, sediment supply, grain size, incision mechanism, and others). Therefore, we assess a wide range of erodibility factors to address uncertainty and to provide order-of-magnitude estimates of the vertical incision rates for comparison against predicted thermal erosion rates. The remaining parameters involved in the mechanical erosion rate estimate are related to the stream power, for which we adopt averages of published (Hoke et al., 2011) valley network flow and morphometric data (which are representative only of large valley network systems with lengths of several 10^3 km and interior channel widths greater than 1 km). Base-line estimates of vertical mechanical erosion rates have

been produced by adopting erodibility factor values of 10^{-7} (for the ice-free substrate) and 10^{-8} (for the ice-cemented substrate) (suggested approximate values; Hurwitz et al., 2010, 2012). The parameter values utilized, and ranges explored, in this assessment of mechanical erosion rates are summarized in Table 1.

For these base-line parameters, we find mechanical incision rates of $\sim 5 \times 10^{-6}$ m/s in the ice-free substrate and $\sim 5 \times 10^{-7}$ m/s in the ice-cemented substrate (Figure 8). While the magnitude of these base-line predicted mechanical erosion rates is substantial (~ 15 - 150 m/yr) (likely related to the high stream power values resulting from the large valley network morphometric parameters adopted; Hoke et al., 2011), they are broadly comparable to observed erosion rates from some terrestrial river systems in active phases of down-cutting and lateral erosion (up to 40 m/yr; Costard et al., 2003; Simon and Rinaldi, 2006). Over a broader range of possible erodibility constants (Figure 8) (Hurwitz et al., 2010, 2012) and stream powers (the range of reported parameter values used to derive the average stream power estimate; Hoke et al., 2011), the mechanical erosion rate spans many orders of magnitude from $\sim 10^{-13}$ to 10^{-1} m/s. While the uncertainties involved in this analysis yield a very substantial range in erosion rate estimates, we will later discuss the insights that are produced.

5. Thermal Erosion Rates

Once the incising fluvial channel has reached the ice-table (Figure 2), vertical mechanical erosion will become more difficult as a result of the increased strength of the underlying material. However, at this point, the underlying ice-cemented substrate is subjected to both mechanical and thermal erosion. The conceptual model assessed here (Figure 2) predicts that mechanical rates in excess of thermal erosion rates will result in a

relative enhancement of lateral erosion rates while lower mechanical erosion rates will have minimal effect on channel formation. In order to evaluate this prediction, valley network mechanical erosion rates were estimated in the previous section, and thermal erosion rates are now assessed.

The rate of thermal erosion is governed by the flux of heat transferred into the substrate by the overlying fluid flow. Here we assume the valley networks to have been formed by fluvial flow in a turbulent regime. This assumption is adopted as the critical Reynolds number for the transition from laminar to turbulent flow over a flat surface (a reasonable approximation for the reaches of most valley networks which exhibit slopes well below $\sim 1^\circ$; *e.g.* Hoke et al., 2011), 3×10^5 (Schlichting, 1955; Aguirre-Puente et al., 1994), is exceeded for fluvial flows with a depth and velocity of only 0.5 and 1 m/s, far below typical estimates for the valley networks (*e.g.* Hoke et al., 2011). In a turbulent flow regime, the heat flux transferred to the substrate by a fluvial flow is given by (Costard et al., 2003):

$$q = h (T_w - T_m) \quad (6)$$

where q (W/m^2) is the induced heat flux, h ($\text{W/m}^2 \text{ K}$) is a heat transfer coefficient dependent upon the Nusselt number, T_w (K) is the water temperature, and T_m is the melting temperature of ice (273 K). A range of plausible valley network water temperatures, T_w , are explored (Figure 8) with a minimum value just above freezing (arbitrarily set to 0.1°C), a nominal value set to the average temperature measured (Von Guerard et al., 1995) from Antarctic melt streams, 3°C , and a maximum value of 60°C

as measured (Edwards et al., 2015) from the meltwater produced by terrestrial volcano-ice interactions (yielding respective temperature differences of 0.1, 3, and 60 K; Figure 8). The heat transfer coefficient h , is governed by the average flow depth, the thermal conductivity of the water, and the flow Nusselt number which is calculated through an empirical relationship dependent upon the Prandtl and Reynolds numbers (Costard et al., 2003). Appropriate Prandtl number and thermal conductivity values are used for each water temperature investigated, and an average flow depth determined from the reported morphometric data is adopted (Hoke et al., 2011). Following (Aguirre-Puente et al., 1994), the Reynolds number is assumed to remain constant at a critical value of 3×10^5 , the critical value at which laminar flow transitions to turbulent flow over a flat surface (a reasonable approximation for the valley network reaches we consider which exhibit slopes far less than 1° and which are likely to transition to turbulent flows over short distances; Aguirre-Puente et al., 1994; Hoke et al., 2011).

Thermal erosion rates of the ice-cemented regolith are estimated through the implementation of a 1-D finite volume heat conduction model (with the phase-change of pore-ice accounted for by the heat integration method; Hu and Argyropoulos, 1996), similar to that applied by Cassanelli and Head (2018). The model domain is constructed to represent a 0.5 km thickness of ice-cemented material in order to evaluate thermal erosion and incision over an average valley network depth of ~ 100 m (Figure 4; Williams and Phillips, 2001) while minimizing potential unphysical temperature increases resulting from the insulating boundary condition that is applied at the model domain base (Figure 9a). Thermal properties of the ice-cemented regolith (*i.e.* conductivity and specific heat capacity) are adopted from (Clifford et al., 2010) while the porosity and density of the

substrate are assumed to follow an exponential decay law (Clifford, 1993) with an updated decay constant of 4.28 km derived from Gravity Recovery and Interior Laboratory (GRAIL; Zuber et al., 2013) mission gravity measurements of the lunar crust (Besserer et al., 2014) and adjusted for martian gravity (Weiss and Head, 2017a, 2017b). The model domain is divided into $\Delta z = 0.05$ m cells on a uniform grid and run forward in time using time steps (Δt) equal to $\sim 1/100$ the von Neumann stability criteria ($\Delta t \leq \Delta z^2 / 2 \kappa$) where the stability criteria is calculated using the maximum thermal diffusivity (κ in m^2/s) value in the model domain to obtain a minimum time step size. The performance of the numerical thermal model using these resolution parameters has been assessed through a benchmarking analysis which is later discussed. At the upper boundary of the model, representing the contact of the ice-cemented material and the fluvial flow, a constant heat flux boundary condition is applied with a value equal to that calculated by equation (3) for each investigated water temperature (Figure 8 and 9). The model assumes that any thawed material is immediately removed from the system, and therefore the thermal erosion rates predicted by this analysis represent maximum estimates (as any accumulated thawed material would act to inhibit the rate of heat transfer from the surface to the underlying ice-cemented material). Models are run for each applied surface heat flux (depending upon the investigated water temperature) until the thickness of melted material reaches the average incised depth of the valley networks, ~ 100 m (Figure 4; Williams and Phillips, 2001), with thermal erosion rates tracked throughout. Relevant physical parameter values utilized, and ranges explored, in the analysis of thermal erosion rates are summarized in Table 1.

To evaluate the suitability of the numerical model and the selected resolutions a benchmark analysis has been performed following the same procedure as Cassanelli and Head (2018). The numerical model is applied to a 1-D one-phase melting Stefan problem for comparison against an analytical solution (the derivation of which is included in Cassanelli and Head, 2018). The one-phase melting Stefan problem used to perform the benchmarking involves the melting of a water-ice half-space with an initially uniform temperature distribution at the melting point to which a constant elevated temperature boundary condition is applied. The benchmark problem is configured here with a Stefan number of ~ 9 , which corresponds to an elevated boundary temperature of 1000 K (and all other relevant parameters set equal to typical values for water). Solutions of the melt-front position versus time for the benchmark problem are generated with the numerical model using a range of model resolutions (Figure 9b) to assess solution convergence and for comparison against the analytical solution to determine accuracy. The benchmark analysis (Figure 9b) indicates close agreement between the numerical model and the analytical solution for the benchmark Stefan problem at the increased resolutions used for our modeling, supporting the validity of the model and the implemented resolution parameters.

Results of the thermal erosion analysis indicate that after the channel base has initially reached the ice-table, melting is delayed by ~ 0.75 , 0.0023 , and 4.9×10^{-5} yr (which is approximately equal to 270 days, 1 day, and 0.5 hr) for fluvial water temperatures in excess of the melting point by 0.1, 3, and 60 K, respectively. These delays in melting are due to the time required to warm the pore-ice to the melting point. Once melting begins, thermal erosion proceeds at an average rate of $\sim 6 \times 10^{-8}$, 1.8×10^{-6} ,

and 2.8×10^{-5} m/s, with total long-term average melting rates including the transient warming period of $\sim 5.8 \times 10^{-8}$, 1.8×10^{-6} , and 2.8×10^{-5} m/s for water temperatures in excess of the melting point by 0.1, 3, and 60 K, respectively (Figure 8).

In general there is a substantial amount of overlap between the predicted mechanical and thermal erosion rates (Figure 8), causing difficulties in precisely distinguishing their relative efficiencies during valley network formation. However, there are several important results produced by this analysis: (1) At low stream power, thermal erosion rates are predicted to exceed mechanical erosion rates (due to efficient heat transfer from the channel water to the substrate) except under limited conditions, requiring high substrate erodibility factors $> \sim 10^{-6} \text{ Pa}^{-1}$ as well as low fluvial water temperatures (Figure 8). (2) For substrate erodibility constants $< \sim 10^{-11} \text{ Pa}^{-1}$, mechanical erosion is predicted to be rate-limiting (the slowest component of substrate erosion) under all relevant valley network conditions (Figure 8). (3) At low meltwater temperatures ($< \sim 3^\circ \text{C}$), mechanical erosion rates can exceed thermal erosion rates under a wider range of conditions. In this case, valley network fluvial channels are predicted by the conceptual model (Figure 2b,1) to undergo preferential lateral expansion at the ice-table, resulting in increased width-to-depth ratios. These low meltwater temperatures ($< \sim 3^\circ \text{C}$) are consistent with a cold and icy early Mars environment (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) and could have been established by: (1) an origin as cool melt derived from surface snow and ice deposits with further heat loss to surface snow/ice encountered during runoff and flow, (2) contact heat loss to a cold, ice-cemented substrate, (3) heat loss to a cold ambient climate during the course of flow. Given the significant integrated lengths of many valley network systems, up to ~ 5000 km

(Fassett and Head, 2008), extended exposure to these cooling conditions is likely, and thus low fluvial water temperatures (with mechanical erosion rates exceeding thermal erosion rates) are expected in the nominal case.

Due to the large uncertainties associated with several critical parameters (*e.g.* the substrate erodibility and the valley network fluvial water temperatures), the quantitative analysis performed here to test the conceptual model (Figure 2) is unable to precisely distinguish the relative efficiency of mechanical versus thermal erosion rates (Figure 8). Therefore, to further evaluate the conceptual model, and to help refine the predictions of the quantitative analysis, we now perform a global assessment of valley network morphometry (Williams and Phillips, 2001).

6. Valley Network Morphometric Analysis

The valley network conceptual formation model predicts preferential lateral erosion and channel/valley expansion when mechanical erosion rates exceeded thermal erosion rates during incision below the ice-table (Figure 2). The quantitative analysis (outlined in Sections 4 and 5) predicts this to occur over a wide range of conditions when valley formation occurs by discharge of water at low temperatures (Figure 8). Consequently, systematically increased valley network width-to-depth ratios and preferentially U-shaped channel/valley cross sections (compared to V-shaped) are predicted to occur in areas of lower ice-table depth, as valleys in these locations would have interacted with the ice-table to a greater extent. In order to test these predictions we utilize a global valley network morphometric data set (Williams and Phillips, 2001) (global data distribution illustrated in Figure 10 showing width-to-depth ratio measurements) to assess potential correlations between predicted ice-table depths and valley network width-to-

depth and occurrence of U-shaped cross sections. Morphometric data utilized from this global catalog (Williams and Phillips, 2001) include valley network widths and depths derived from individual Mars Orbiter Laser Altimeter (MOLA; Smith et al., 2001) shot point topographic measurements (therefore not subject to interpolation errors) taken from available MOLA track crossings of valley network systems.

Results of this global morphometric analysis reveal a weak tendency toward increased width-to-depth ratios at shallower ice-table depths, consistent with the conceptual model predictions (Figure 11a). There is no significant correlation (Figure 11a), however, between valley network width-to-depth ratios and estimated ice-table depth, with a large degree of variance in the data resulting in a poor linear regression correlation coefficient and a large spread in the confidence bounds. As an additional test, the correlation between valley network width-to-depth and elevation has been assessed (Figure 11b). This test has been performed because under the cold and icy conditions that climate models predict to characterize the early Mars environment, adiabatic cooling results in decreasing temperatures with increasing elevation. Equilibrium ice-table depth exhibits a direct dependence on the mean annual surface temperature as is illustrated by the global ice-table distribution map (Figure 5). Therefore, evaluating the correlation between valley network width-to-depth ratios and elevation instead of the ice-table depths offers a test which damps the potential influence of other factors that might alter ice-table depths such as variations in substrate thermal conductivity and geothermal heat flux which we have assumed to be constant across the surface of Mars. Results of the correlation assessment between valley network width-to-depth ratios and elevation (Figure 11b), however, reveal a similarly poor correlation as with ice-table depths.

Investigation of the distribution of valley networks exhibiting U-shaped versus V-shaped cross sections (Figure 12) indicates that while valleys characterized by U-shaped cross sections generally exhibit greater width-to-depth ratios (as is expected) (Figure 12a), they also occur at greater ice-table depths than those with V-shaped cross sections (with greater ice-table depths for both the mean and interquartile range bounds) (Figure 12b). While the difference in the ice-table depth distributions between the V-shaped and U-shaped valleys is minimal, this observation contradicts the predictions of the valley network formation conceptual model. The slightly greater ice-table depths observed in association with the U-shaped valleys can potentially be attributed to the fact that U-shaped valleys tend to be more mature systems occurring at lower elevations (with warmer temperatures and greater ice-table depths) (Figure 12c) with greater depths (Figure 12d) and higher stream orders (Figure 12e).

A potential issue related to these morphometric analyses is the reliance on the valley network width-to-depth ratio, a parameter which does not uniquely differentiate between U-shaped and V-shaped valley networks (as evidenced by their similar width-to-depth distributions Figure 12a). In order to address this possible source of uncertainty the former regression analysis of predicted ice-table depth versus valley network width-to-depth ratios has been repeated using exclusively U-shaped valley networks (Figure 13a). In addition, a regression assessment of predicted ice-table depth versus U-shaped valley network bottom width has been performed (Figure 13b) to isolate the variance introduced by the width-to-depth parameter (as is discussed in the following section). These additional regression analyses, however, again show no evidence for a correlation

between the predicted ice-table depths and U-shaped valley network morphometric parameters.

In summary, the results of the morphometric analysis suggest that either: (1) the permafrost ice-table did not significantly influence valley network width-to-depth ratios or morphology, possibly owing to more rapid thermal erosion rates than mechanical, (2) permafrost was not present in the Noachian Mars substrate, indicating a warmer climate, or (3) that additional factors not considered in this analysis are involved (reviewed in more detail in the following section) which may have altered ice-table depths or valley network width-to-depth ratios and morphology. We now review a collection of possible additional factors not considered in this analysis and their potential influence on our findings.

6.1 Valley Network Morphometric Analysis: Complicating Factors

There are a number of complicating factors which may have altered both the predicted equilibrium ice-table depths and valley network morphometric parameters which have not been accounted for in our analysis.

Factors that may have influenced the predicted equilibrium ice-table depths include:

(1) Geothermal heat flux: In our analysis we adopted a constant global average geothermal heat flux of 55 mW/m^2 . This is a clear oversimplification as the surface geothermal heat flux of Mars is likely to vary considerably across the surface (Plesa et al., 2016), particularly during the Noachian period with the widespread occurrence of volcanic activity (Carr and Head, 2010; Cassanelli et al., 2015; Cassanelli and Head, 2016). (2) Substrate thermal conductivity: While we evaluated a wide possible range of substrate thermal conductivities, each tested value was assumed to be constant across the

surface of Mars. Again this is a clear oversimplification, with thermal conductivity of the substrate likely to vary substantially (potentially by several orders of magnitude) across the surface of Mars for different surface materials and physical states. (3) Prevailing climate conditions: Here we adopted a typical Late Noachian early Mars climate scenario characterized by a 1 bar CO₂ atmosphere and 45° planetary obliquity (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018). However, variations in the atmospheric state and orbital parameters would influence the distribution and magnitude of mean annual surface temperatures and water abundance (*e.g.* Figure 3), therefore effecting predicted equilibrium ice-table depths.

Factors which may have influenced valley network morphometric parameters include:

(1) Substrate heterogeneity: Variations in the composition and physical state (*e.g.* porosity) of the substrate across Mars, not accounted for here, would have resulted in variable rates of mechanical and thermal erosion (due to variations in substrate strength and ice content). Therefore during valley network incision below the ice-table, variations in the relative efficiency of mechanical erosion (above and below that of thermal erosion) could have led to variations in the relative efficiency of lateral erosion and, by consequence, valley network width-to-depth ratios. (2) Valley preservation state: The valley network width-to-depth ratios used to perform the morphometric analysis were measured from preserved fluvial valleys and not from the original channels (see Carr, 1996, p. 47-100 for discussion of the importance of this distinction). This is a simplification adopted out of necessity, as there are very few instances of preserved valley network fluvial channels on Mars. The use of valley width-to-depth ratios instead of channels thus incorporates the influence of other processes and factors which have

modified the post-formation valley profiles. These include the strength/cohesion of the substrate materials in which the valley is formed, the angle of repose of the substrate material, mass wasting processes, aeolian erosion and infilling, and modification by glacial processes (*i.e.* lineated valley fill; Head et al., 2006a, 2006b, 2010). These factors will have the effect of generally increasing the post-formation valley network width-to-depth ratios. (3) Sampling bias: The global data set (Williams and Phillips, 2001) used to perform the morphometric analysis contained ~2,000 measurements of valley networks, with measurement points dictated by locations where MOLA crossings of valley network systems occurred at the time of the study. Therefore, an updated, more complete and consistent sampling of valley network morphometry might yield an improved correlation between ice-table depth and width-to-depth ratios. (4) Position of measurement within valley network system: Here we performed an indiscriminate evaluation of a complete global data set (Williams and Phillips, 2001) of valley network morphometric measurements. However, individual valley network systems are likely to experience variations in substrate and flow conditions between and along the tributary and trunk segments. These variations, and their potential influence on valley morphometric parameters, may also be compounded further when examining different valley network systems. Examples of possible variations include changes in substrate erosive resistance (*e.g.* loose regolith versus consolidated material, with increasing erosive resistance resulting in decreased mechanical erosion rates) and valley floor slope (with increased slope resulting in increases in both mechanical and thermal erosion rates). Therefore, potential remains for trends to be identified between ice-table depth and valley network morphometry through more focused future analyses evaluating: (1) comparable sections

of the same or different valley network systems which may have experienced comparable substrate and flow conditions, and (2) individual valley network systems as a function of distance along the system courses.

All of the above-mentioned factors are likely to contribute to variations in both predicted equilibrium ice-table depths and valley network morphometric parameters which would influence any potential correlations. Therefore, these factors represent areas for future investigation to refine the predictions and analyses performed in this work.

7. Conclusions

We performed quantitative and morphometric analyses to evaluate a conceptual model for the formation of valley networks in a cold and icy early Mars climate in order to determine if the valley network characteristics are consistent with a cold formative environment.

Results of the quantitative analysis indicate that, under cold and icy conditions, the global ice-cemented substrate (permafrost) occurs below a ~50-100 m thick layer of ice-free regolith at the surface, which forms the ice-table at the interface with the underlying ice-cemented material. In the valley network regions, the depth to the ice-table also ranges from ~50-100 m with an average of ~80 m. Given that the depths of most valley network systems exceed those estimated for the ice-table, interactions between incising valley networks and the ice-table are found to have been possible. In addition, (1) the estimated depths of the Noachian ice-table indicate that meltwater could infiltrate and mechanically erode the substrate for an extended period before reaching the ice-table (freezing of infiltrating meltwater could also have an influence), and (2) that any warm periods would have to be of extended duration (~2 kyr) in order for the surface

temperatures to penetrate to the ice-table to cause extensive melting of ground ice to induce solifluction/gelifluction activity. The estimated ice-table depths on the order of tens of meters lead to formation of a dry, ice-free surface regolith which could efficiently absorb any precipitation generated by transient atmospheric heating events, potentially inhibiting or preventing surface runoff. Assessment of the potential infiltration capacity of the martian regolith, however, indicates that surface runoff could be readily generated at plausible rainfall rates ($\sim 1\text{-}10\text{ mm/hr}$).

Evaluation of the relative rates of mechanical erosion of the substrate and thermal erosion rates of the ice-cement during valley network incision below the ice-table are subject to large uncertainties in a number of critical parameters (*i.e.* substrate erodibility and fluvial water temperature). However, the general findings yield several important results. (1) At low stream power, thermal erosion rates are predicted to exceed mechanical erosion rates (due to efficient heat transfer from the channel water to the substrate) except under limited conditions, requiring high substrate erodibility factors $>\sim 10^{-6}\text{ Pa}^{-1}$ as well as low fluvial water temperatures. (2) For substrate erodibility constants $<\sim 10^{-11}\text{ Pa}^{-1}$, mechanical erosion is predicted to be rate-limiting under all relevant valley network conditions. (3) At low meltwater temperatures, consistent with a cold and icy early Mars environment, mechanical erosion rates can exceed thermal erosion rates under a wide range of conditions. In this scenario, channels are predicted by the conceptual model to undergo preferential lateral expansion at the ice-table, resulting in increased width-to-depth ratios.

To improve upon the results of the quantitative analysis, a global morphometric analysis of the valley networks was performed to assess any correlations between

predicted ice-table depths and valley network width-to-depth ratios, bottom widths, and occurrence of U-shaped cross sections. This global morphometric analysis revealed a weak tendency toward increased valley network width-to-depth ratios at shallower ice-table depths, consistent with the conceptual model predictions. No statistically significant correlation was found, however, between the investigated valley network morphometric parameters and estimated ice-table depth, suggesting that either: (1) the permafrost ice-table did not significantly influence valley network width-to-depth ratios, possibly owing to more rapid thermal erosion rates than mechanical, (2) permafrost was not present in the Noachian Mars substrate, indicating a warmer climate, or (3) that additional factors are involved. Potential additional factors which may have influenced the results of this correlation include: variability in the surface geothermal heat flux, substrate thermal conductivity, substrate composition and physical state, as well as the valley network preservation state, and data set sampling bias. These factors represent areas for future investigation and improvement in order to more precisely evaluate the formation of valley networks in a cold early Mars climate.

Acknowledgements

The quality and clarity of this manuscript was improved by thorough and insightful comments from two anonymous reviewers. We thank David Weiss, Marc Parmentier, Jim Russell, and Caleb Fassett for helpful comments and discussion. Thanks are extended to Rebecca M. E. Williams and Ashley Palumbo for sharing data published in Williams and Phillips (2001) and Palumbo et al. (2018). We gratefully acknowledge support to JWH for participation in the Mars Express High-Resolution Stereo Camera Team (JPL 1237163).

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<https://doi.org/10.1126/science.1231507>

Parameter	Symbol	Principal Variation Conditions	Range Evaluated	Nominal Value	Units
Latent heat of fusion, ice	L	Pressure	--	3.35×10^5	J/kg
Saturation vapor density of water	n_s	Temperature; <i>computed see Section 2</i>	--	--	molec./m ³
Mean atmospheric near-surface vapor density	n_{atm}	Temperature; <i>computed see Section 2</i>	--	--	molec./m ³
Atmospheric scale height	H	Atmospheric composition and temperature	--	11	km
Water molecule mass	m	N/A	--	2.989×10^{-26}	kg
Mean annual surface temperature	T	Latitude, atmospheric pressure; <i>see Figure 3</i>	--	--	K
Substrate thermal conductivity	K_{Ts}	Substrate composition and density	0.1-2	0.5	W/m K
Geothermal heat flux	g_h	Subsurface heat production and flow	--	55	mW/m ²
Saturated hydraulic conductivity	K_h	Substrate permeability and fluid density, viscosity	9×10^{-7} to 3×10^{-5}	3×10^{-7}	m/s
Capillary head	ψ	Substrate pore-size and water content	0.04-0.31	0.20	m
Unsaturated substrate porosity	$\Delta\theta$	Substrate porosity and water content	0.3-0.4	0.36	
Martian gravity	g	N/A	--	3.71	m ² /s
Water density	ρ_w	Temperature	--	1000	kg/m ³
Dynamic viscosity of water	μ_w	Temperature	--	1.79×10^{-3}	Pa s
Substrate erodibility factor	K_s	Many, see Sklar and Dietrich (2001) for details	10^{-12} to 10^{-4}	10^{-8} and 10^{-7}	Pa ⁻¹
Average volume flux per channel width	Q	Flow velocity and depth	0.1-10	10	m ² /s
Channel slope	α	Topography	0.1-2	0.09	°
Melting temperature of ice	T_m	Pressure, purity	--	273	K
Valley network water temperature	T_w	Water source	273.1-333	276	K
Valley network water depth	d	Channel size	--	10	m
Valley network flow velocity	u	Channel size, slope, roughness	--	1	m/s
Valley network Prandtl number	Pr	Fluid temperature, composition	3-13	12.4	
Valley network Reynolds number	Re	Flow depth, velocity, viscosity	--	3×10^5	
Valley network Nusselt number	Nu	Flow depth, velocity, viscosity	6×10^3 to 10^4	9×10^4	
Valley network water thermal conductivity	K_{Tw}	Temperature	0.56-0.65	0.56	W/m K
Ice-cemented regolith thermal conductivity	K_{Tr}	Porosity, temperature; <i>computed see Section 2</i>	--	--	W/m K
Ice-cemented regolith specific heat capacity	C_{pr}	Porosity, temperature; <i>computed see Section 2</i>	--	--	J/kg K
Ice density	ρ_i	Porosity, temperature	--	920	kg/m ³
Substrate porosity	ϕ	Subsurface compaction; <i>computed see Section 2</i>	--	--	
Substrate density	ρ_s	Porosity, composition; <i>computed see Section 2</i>	--	--	kg/m ³
Substrate porosity decay constant	K	Gravity	--	4.28	km
Substrate surface porosity	ϕ_o	Crustal degradation state	--	0.3	

Table 1. Summary table of major parameters utilized throughout the presented analyses and, where applicable, the corresponding parameter ranges explored (-- indicates parameter values and ranges not applicable or not considered).

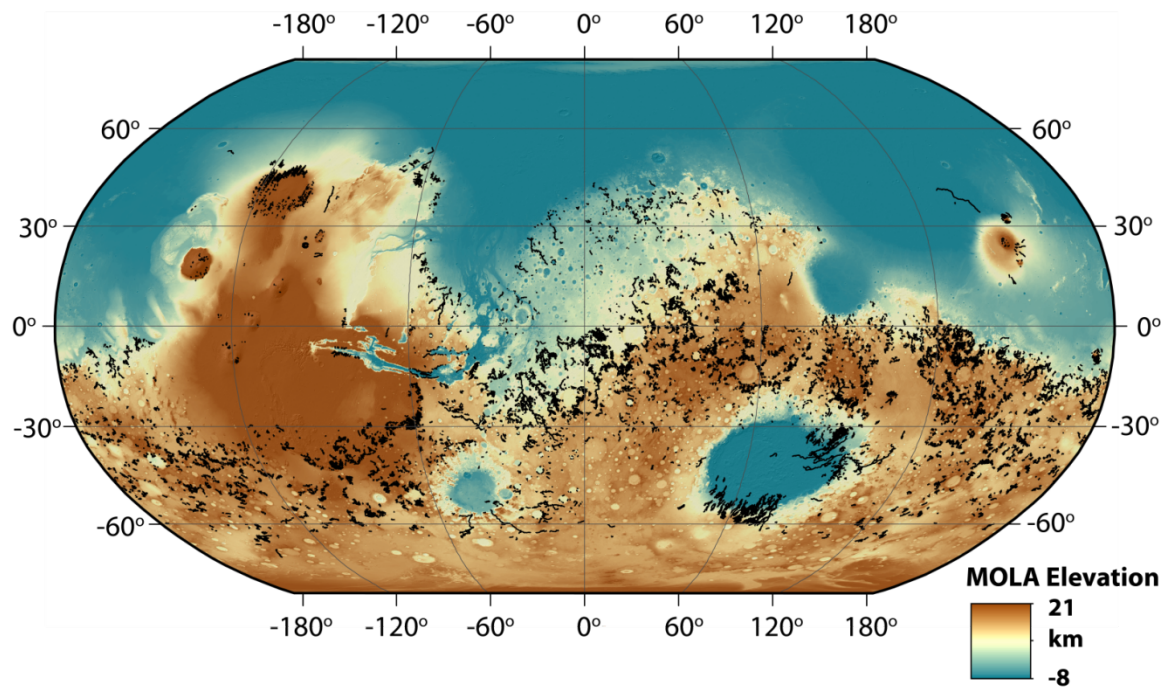


Figure 1. The global distribution of valley networks on Mars (Hynek et al., 2010) with individual valley networks mapped as solid black lines. Base data is comprised of shaded MOLA (Mars Orbiter Laser Altimeter; Smith et al., 2001) topography, overlain by stretched MOLA elevation data.

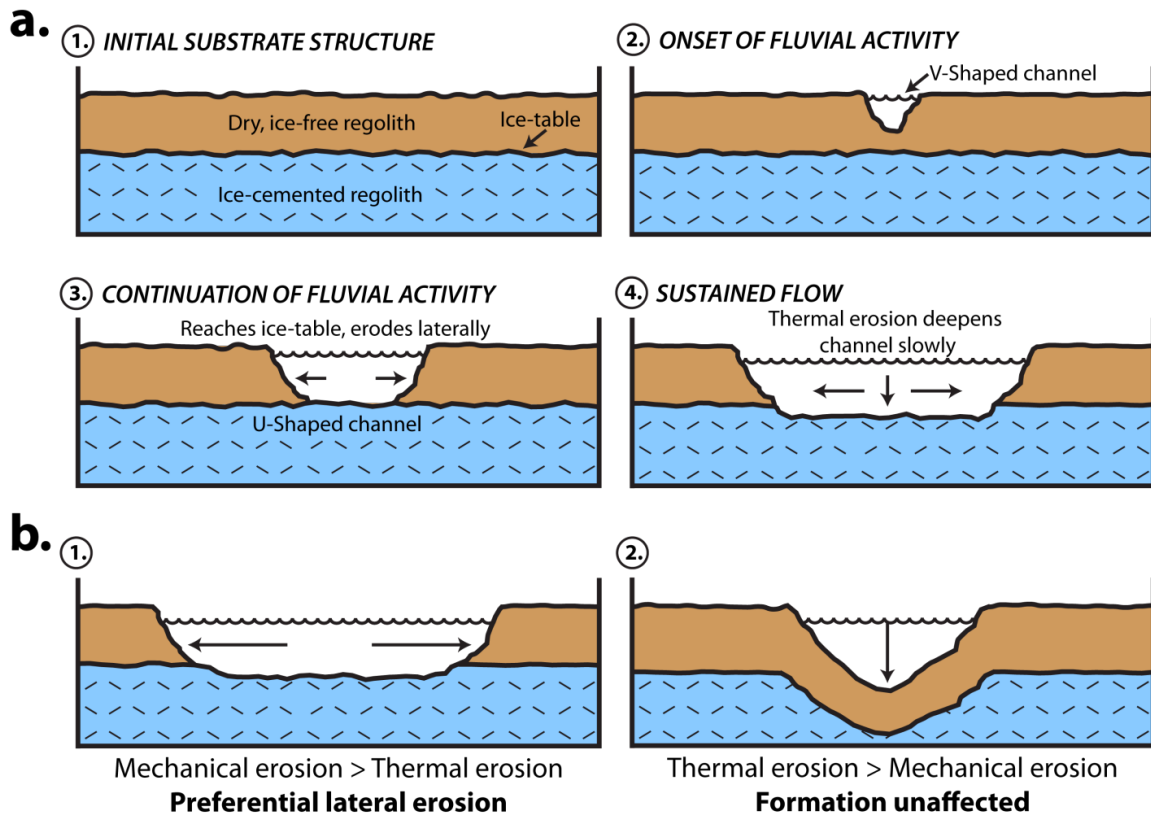


Figure 2. (a.) Conceptual model for valley network fluvial channel formation and evolution under cold and icy conditions (adapted from Head and Cassanelli, 2014). Under cold and icy conditions the fluvial substrate (1) is comprised of an ice-free, dry-active layer and underlying ice-cemented regolith, with the interface between them being the ice-table. During the onset of fluvial activity (2) a V-shaped channel incises the ice-free dry-active layer, which after reaching the ice-table (3), experiences reduced incision rates due to the coherent ice-cemented material and evolves to a flat-floored U-shaped channel. As fluvial flow continues (4), channel incision proceeds at a reduced rate by thermal erosion of the ice-cemented material while lateral erosion continues. (b.) The effect of the relative efficiency of mechanical versus thermal erosion on stream morphology. (1) When mechanical erosion rates of the substrate exceed thermal erosion rates of the ice-cemented the relative efficiency of lateral erosion is predicted to increase, resulting in channels characterized by larger width-to-depth ratios and U-shaped cross section morphologies. (2) In contrast, when thermal erosion rates of the ice-cement are greater than mechanical erosion rates, the fluvial channel is not predicted to directly interact with the ice-table and therefore formation proceeds unaffected.

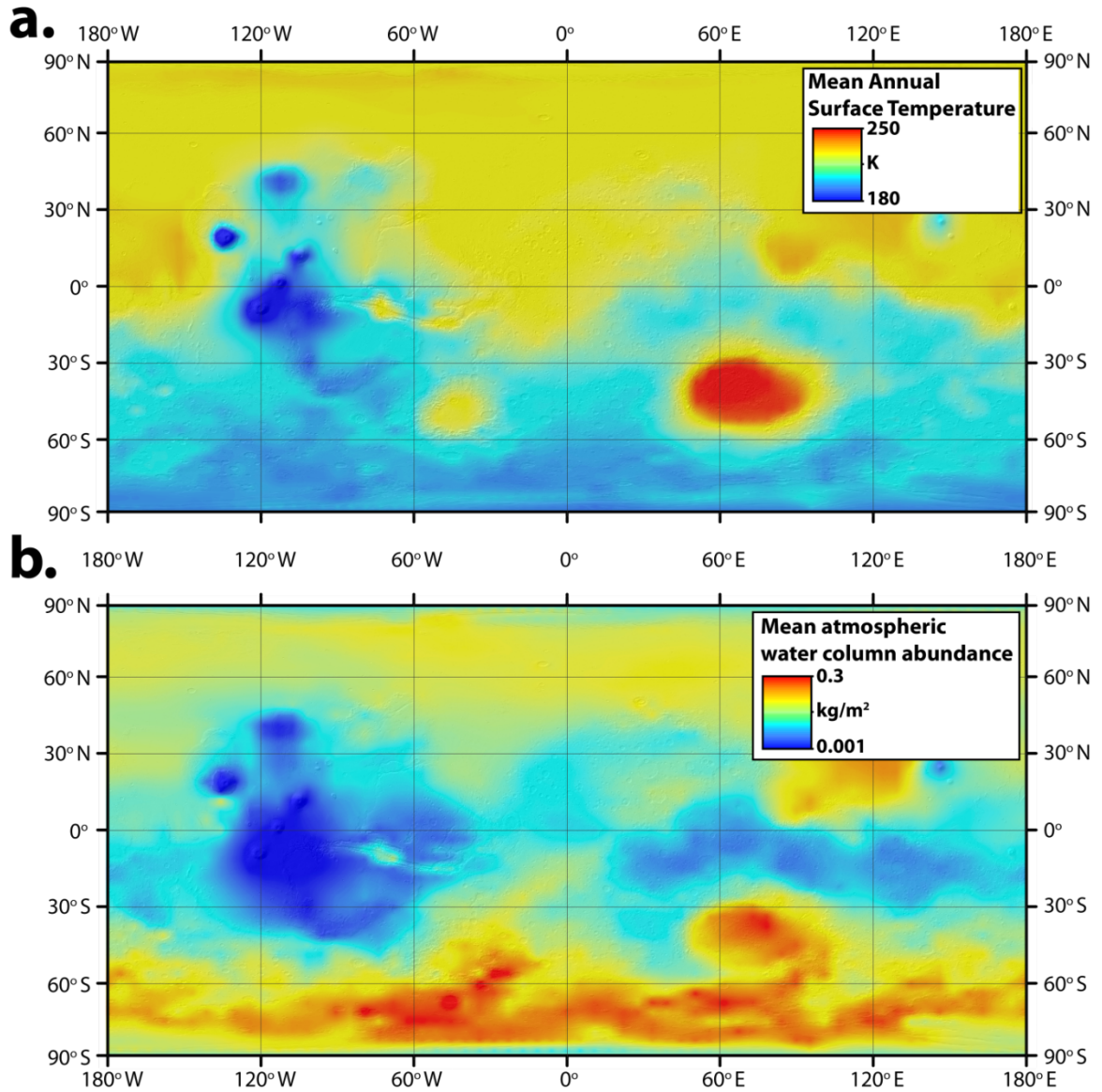


Figure 3. (a.) Map of global mean annual surface temperatures and (b.) mean annual atmospheric water column abundance utilized to estimate Late Noachian equilibrium ice-table depths. Map data are derived from 3D global climate model simulations of a typical Late Noachian Early Mars climate scenario with a 1 bar CO₂ atmosphere and 45° obliquity (Palumbo et al., 2018). Note the strong altitudinal dependence of mean annual surface temperatures (a.) due to the presence of the thicker CO₂ atmosphere and the resulting adiabatic cooling of elevated regions (Forget et al., 2013; Wordsworth et al., 2013).

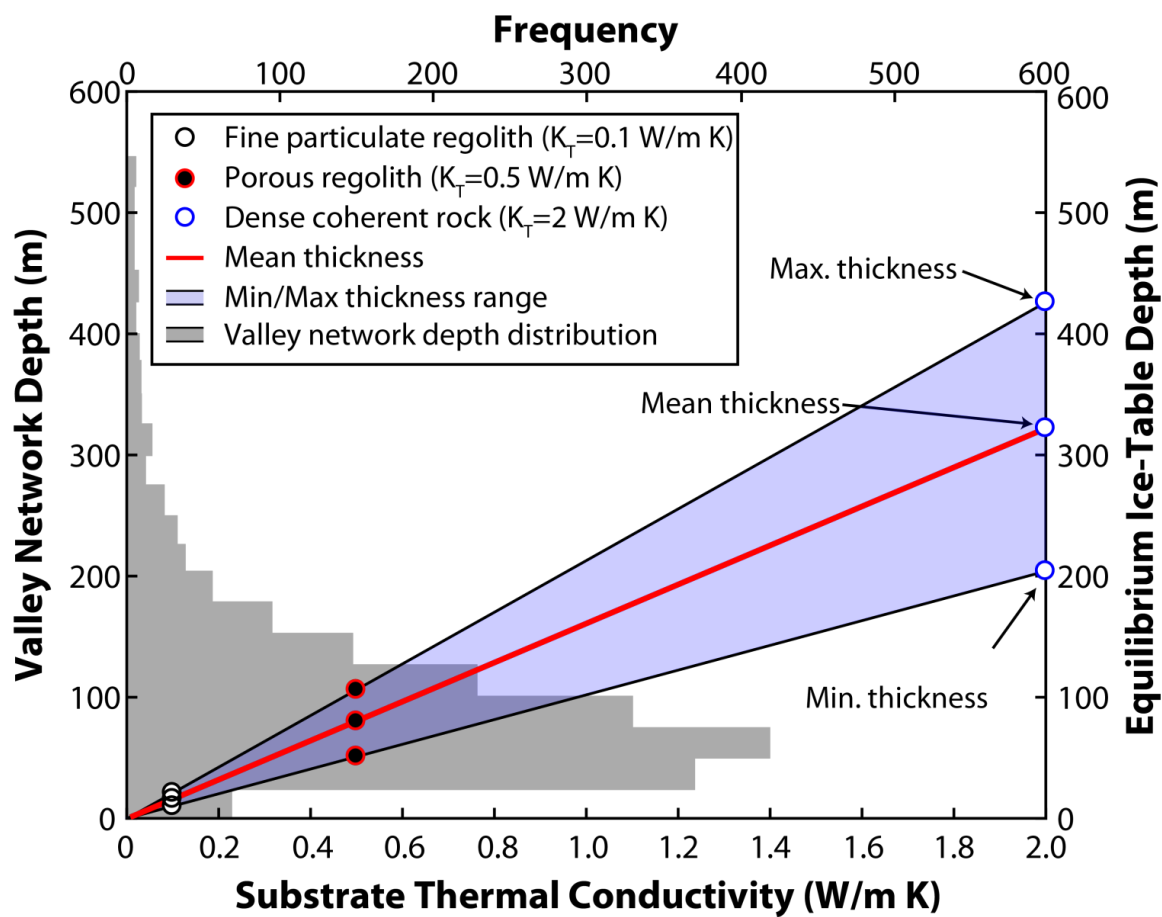


Figure 4. The range of equilibrium ice-table depths along the valley networks (illustrated in Figure 5b) as a function of substrate thermal conductivity plotted (on the right-side y-axis) against the valley network depth frequency distribution (Williams and Phillips, 2001) (shown by the gray histogram on the left-side y-axis). With a nominal substrate thermal conductivity of 0.5 W/m K (representing a porous regolith; Mellon et al., 1997) (Table 1), predicted equilibrium ice-table depths, ~50-100 m, are exceeded by the incised depths of most valley networks, suggesting channel formation could have been influenced by the presence of ice-cemented substrate material.

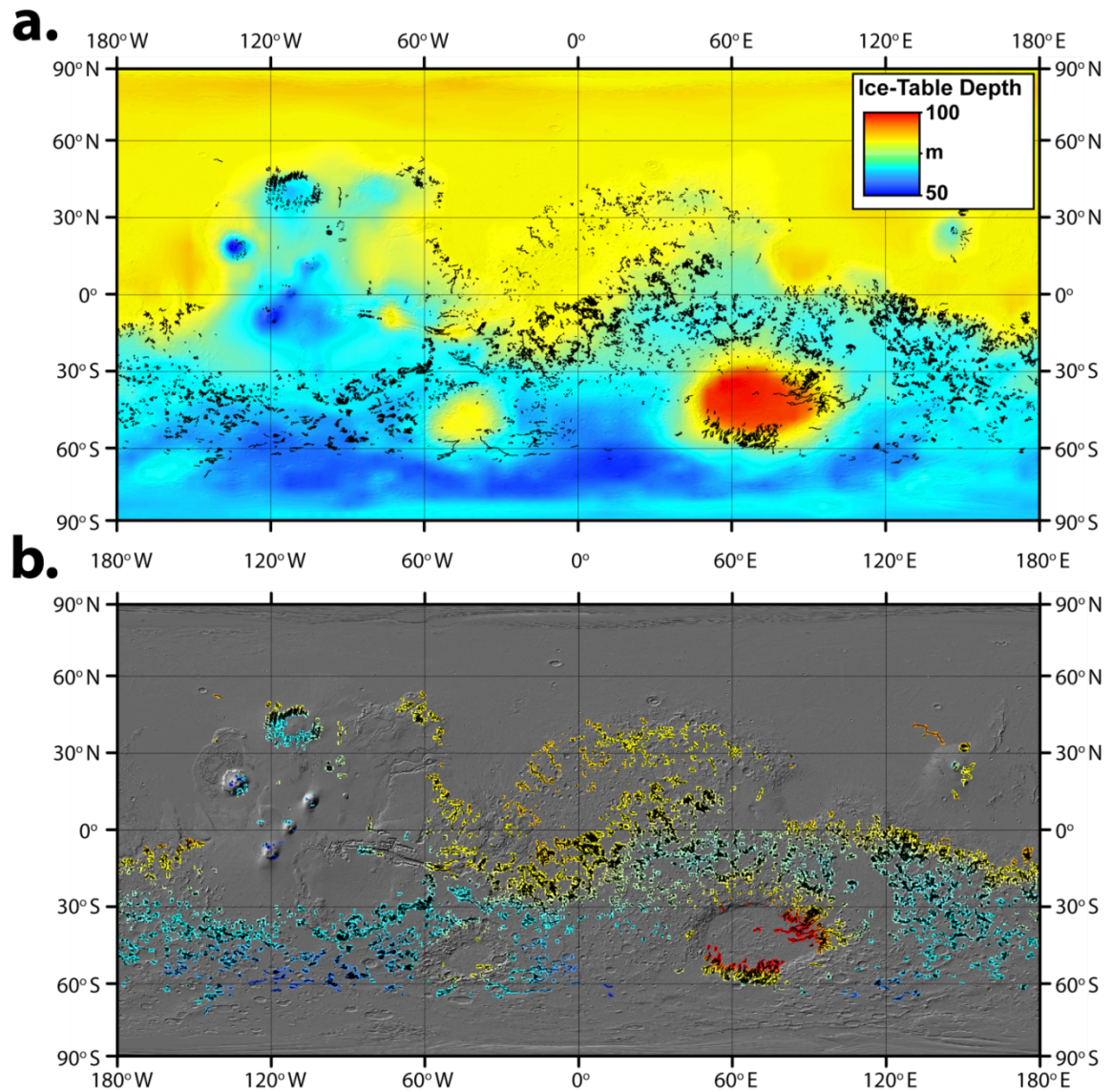


Figure 5. (a.) Global map of equilibrium ice-table depths produced assuming a nominal porous regolith substrate (with a thermal conductivity of 0.5 W/m K; Mellon et al., 1997) (Table 1). Mapped valley networks (Hynek et al., 2010) are depicted by solid black lines. **(b.)** Equilibrium ice-table depths along the valley networks of Mars, ranging from ~50 to 100 m in depth with a mean of ~80 m. Note that ice-table depths decrease with elevation resulting from adiabatically decreasing mean annual surface temperatures in a cold and icy early Mars climate with a thicker CO₂ atmosphere (Forget et al., 2013; Wordsworth et al., 2013). As a result, ice-table depths are generally greater in the warmer northern lowlands and shallower in the colder highland regions.

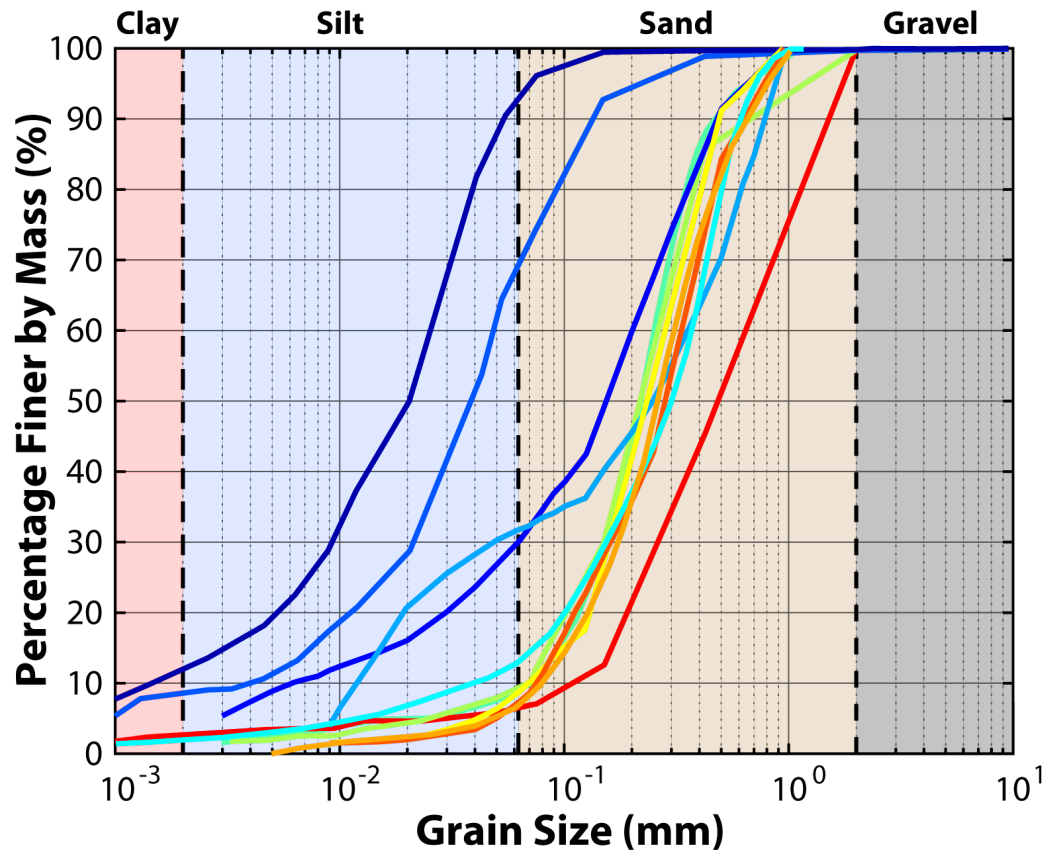


Figure 6. Grain size distributions for a range of Mars regolith simulants with different colors representing different regolith simulants (Skaggs et al., 2001; Seiferlin et al., 2008; Zeng et al., 2015). The simulants are predominantly composed of silt and sand-sized particles with a small clay fraction, similar to a terrestrial clay loam, which is adopted as a representative physical analogue to the Noachian martian regolith.

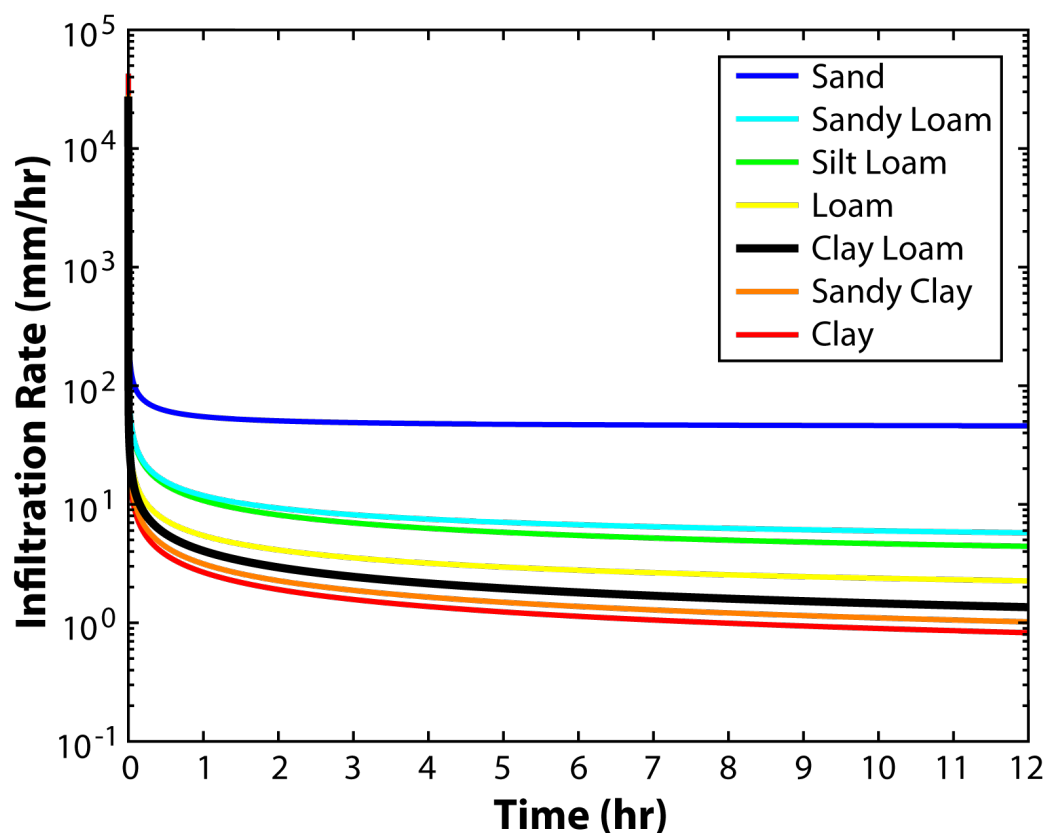


Figure 7. Potential infiltration rates versus time for a range of porous media exhibiting clay to sand dominated grain size distributions. Infiltration rates are calculated using the Green-Ampt method and relevant hydraulic properties (Chow et al., 1988) (Table 1). Results for the clay loam soil, which grain size distributions suggest best represents the martian regolith, are shown by the heavy black line. Over a plausible precipitation event timescale of 12 hr, infiltration rates of the clay loam soil rapidly decay to ~1-10 mm/hr.

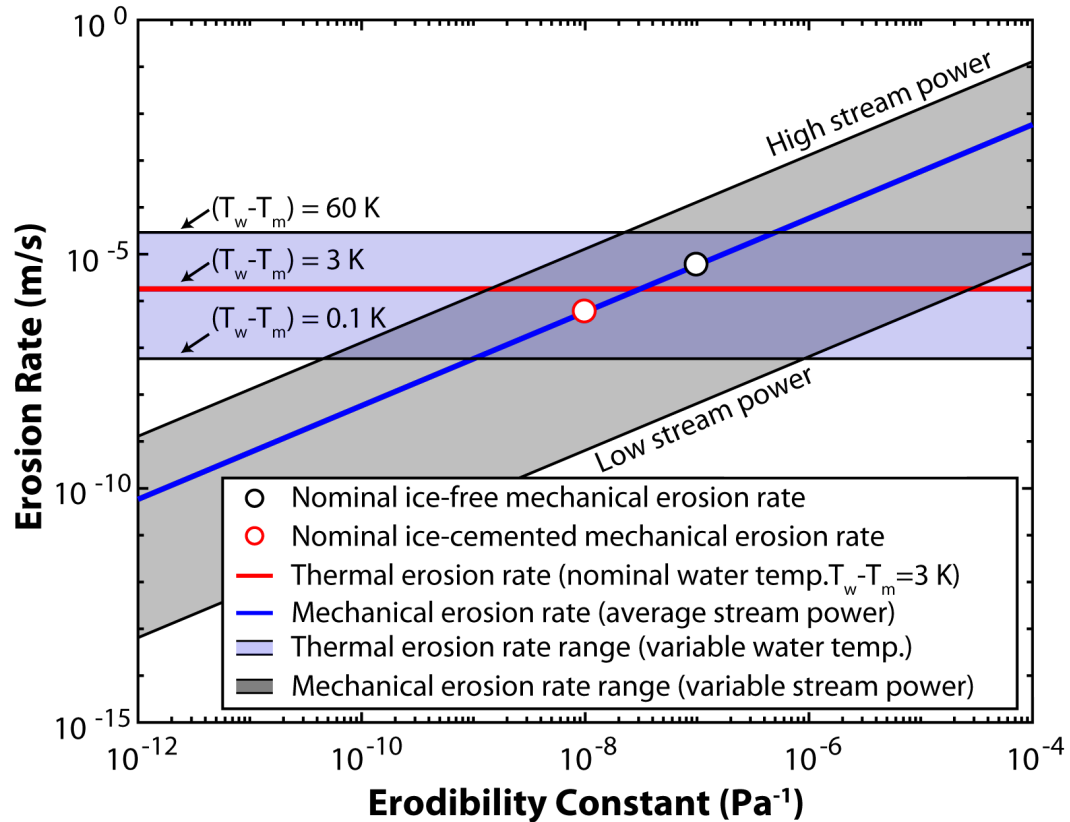


Figure 8. Comparison of predicted mechanical and average thermal erosion rates. A range of mechanical erosion rates are represented by the gray shaded area, corresponding to variations in valley network stream power from published valley network water flux estimates and bed slopes (Hoke et al., 2011). The blue line represents the mechanical erosion rate versus the erodibility of the substrate for the average stream power, with circles indicating the mechanical erosion rates for the adopted substrate erodibility values (10^{-7} and 10^{-8} Pa⁻¹ for the ice-free and ice-cemented substrates respectively). Average thermal erosion rates are shown by the blue shaded area for a range of temperature contrasts between the valley network fluvial water and the ice-melting temperature. The red line indicates a nominal thermal erosion rate corresponding to a fluvial water temperature of 3 °C, an average value measured from Antarctic melt streams (Von Guerard et al., 1995).

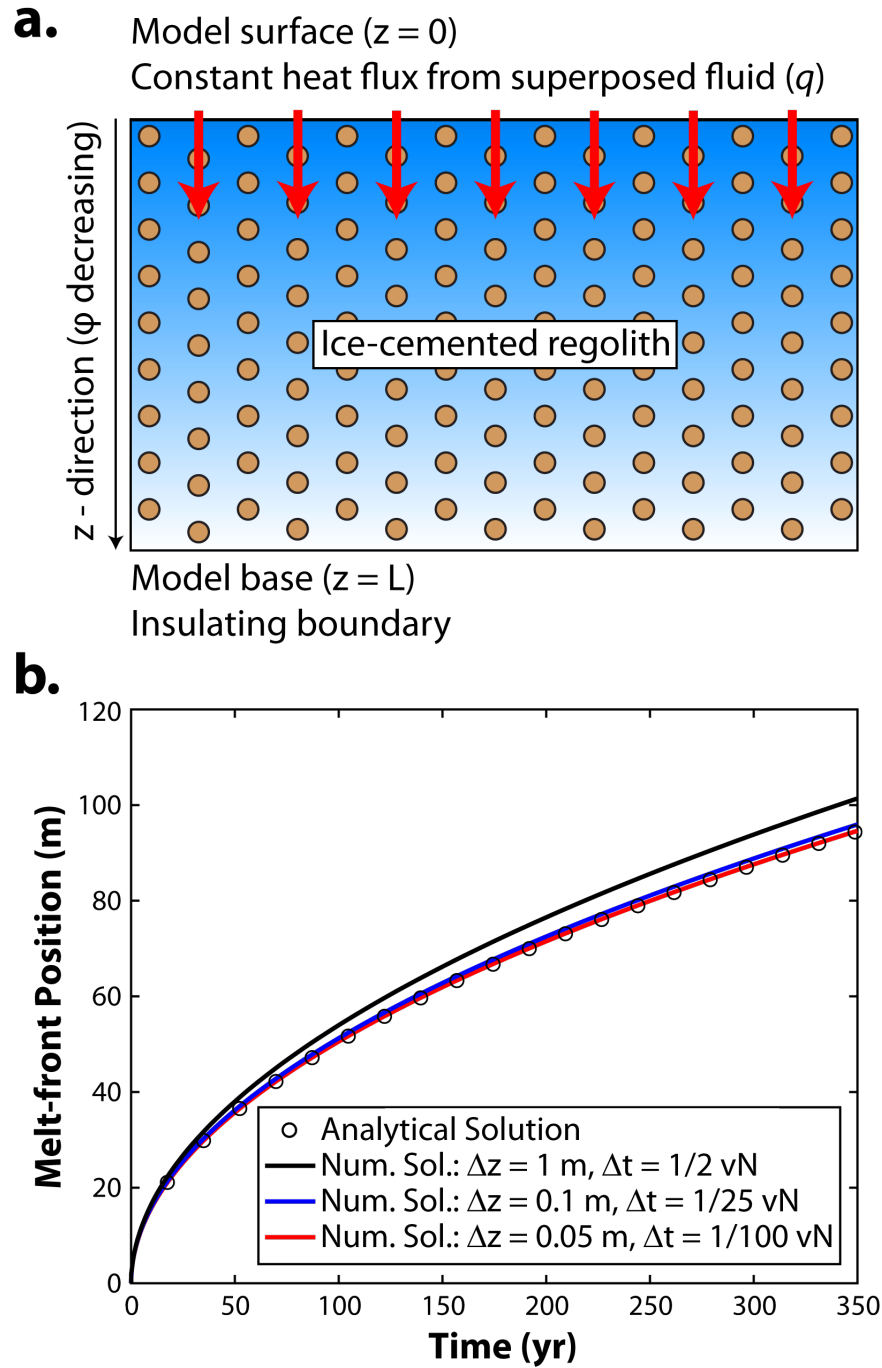


Figure 9. (a.) 2-D schematic diagram illustrating the configuration of the 1-D numerical thermal model used to assess thermal erosion rates within the ice-cemented regolith. Fluid flowing over the surface induces a downward heat flux (q) into the ice-cemented regolith, which exhibits exponentially decreasing porosity (ϕ) with increasing depth (z). The model is constructed with an initial thickness (L) of 500 m to evaluate thermal erosion over an average valley network depth of ~ 100 m while minimizing the effect of the insulating basal boundary condition. Any cells that undergo melting are instantaneously removed from the system, and therefore the resulting thermal erosion

rates represent maximum estimates. **(b.)** Performance of the numerical model on the benchmark one-phase Stefan problem as discussed in Section 5 over a range of spatial ($\Delta z = 1, 0.1, \text{ and } 0.05 \text{ m}$) and temporal model resolutions ($\Delta t = 1/2, 1/25, \text{ and } 1/100$ the minimum von Neumann stability criteria). Numerical predictions of the melt-front position versus time converge to very close agreement with the analytical solution at increased spatial and temporal resolution supporting the validity of the numerical model and the implemented resolution parameters.

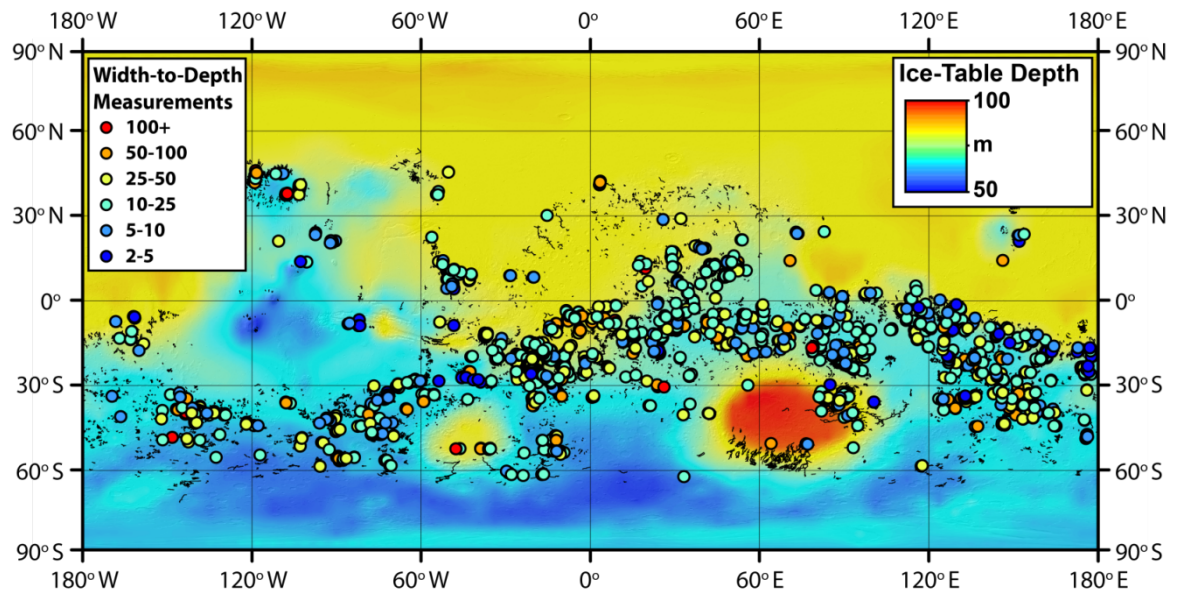


Figure 10. Measurements (Williams and Phillips, 2001) of valley network width-to-depth ratios overlain on a global map of predicted equilibrium ice-table depths (from Figure 5). Spatial correlations between the predicted equilibrium ice-table depths and valley network width-to-depth ratios have been analyzed with results shown in Figure 11.

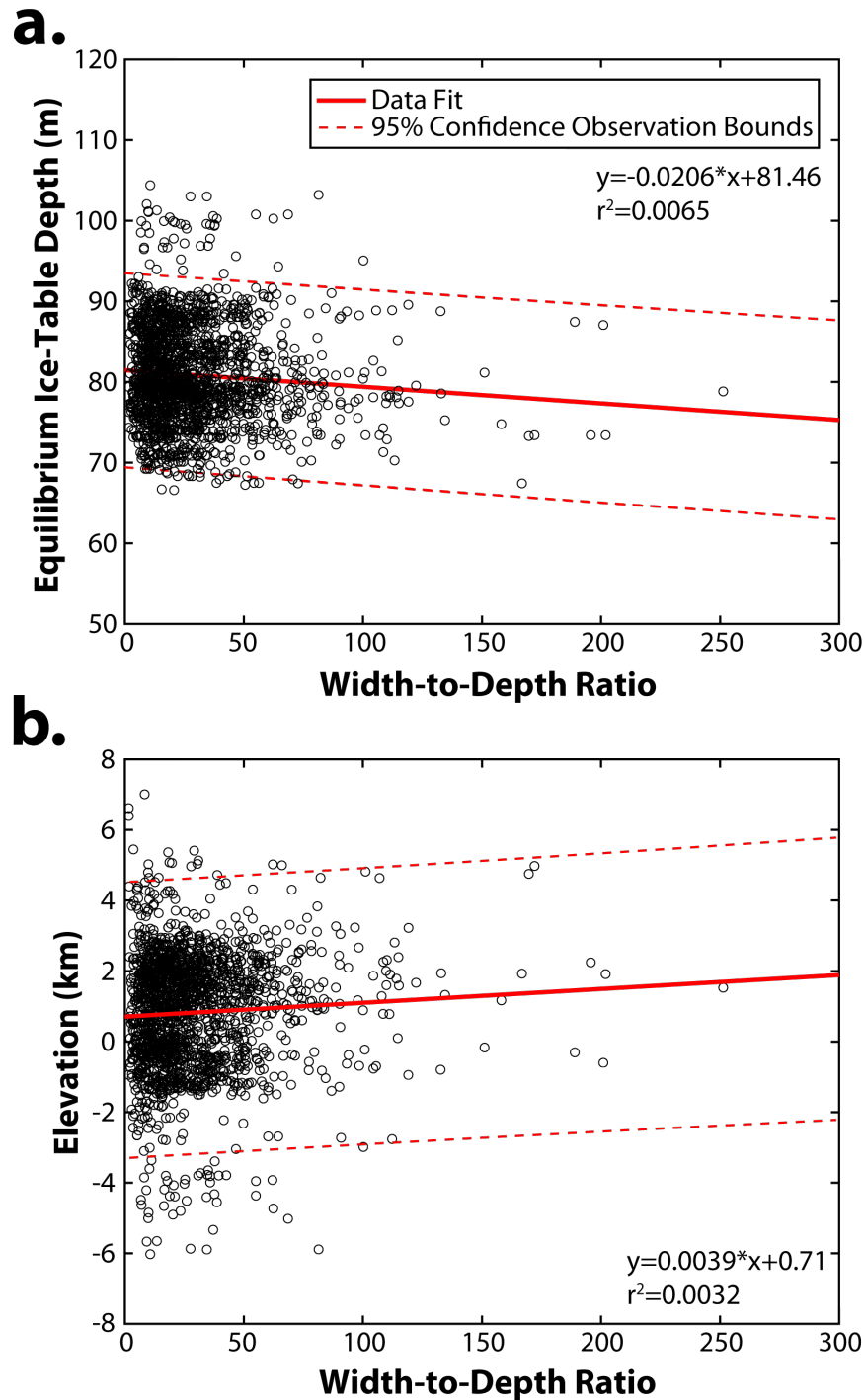


Figure 11. (a.) Measured (Williams and Phillips, 2001) valley network width-to-depth ratios versus predicted equilibrium ice-table depths. **(b.)** Measured (Williams and Phillips, 2001) valley network width-to-depth ratios versus elevation. In both panels the red line represents a linear regression fit and the dashed lines the 95% confidence observation bounds. No significant correlations are found between valley network width-to-depth ratios and ice-table depth or elevation.

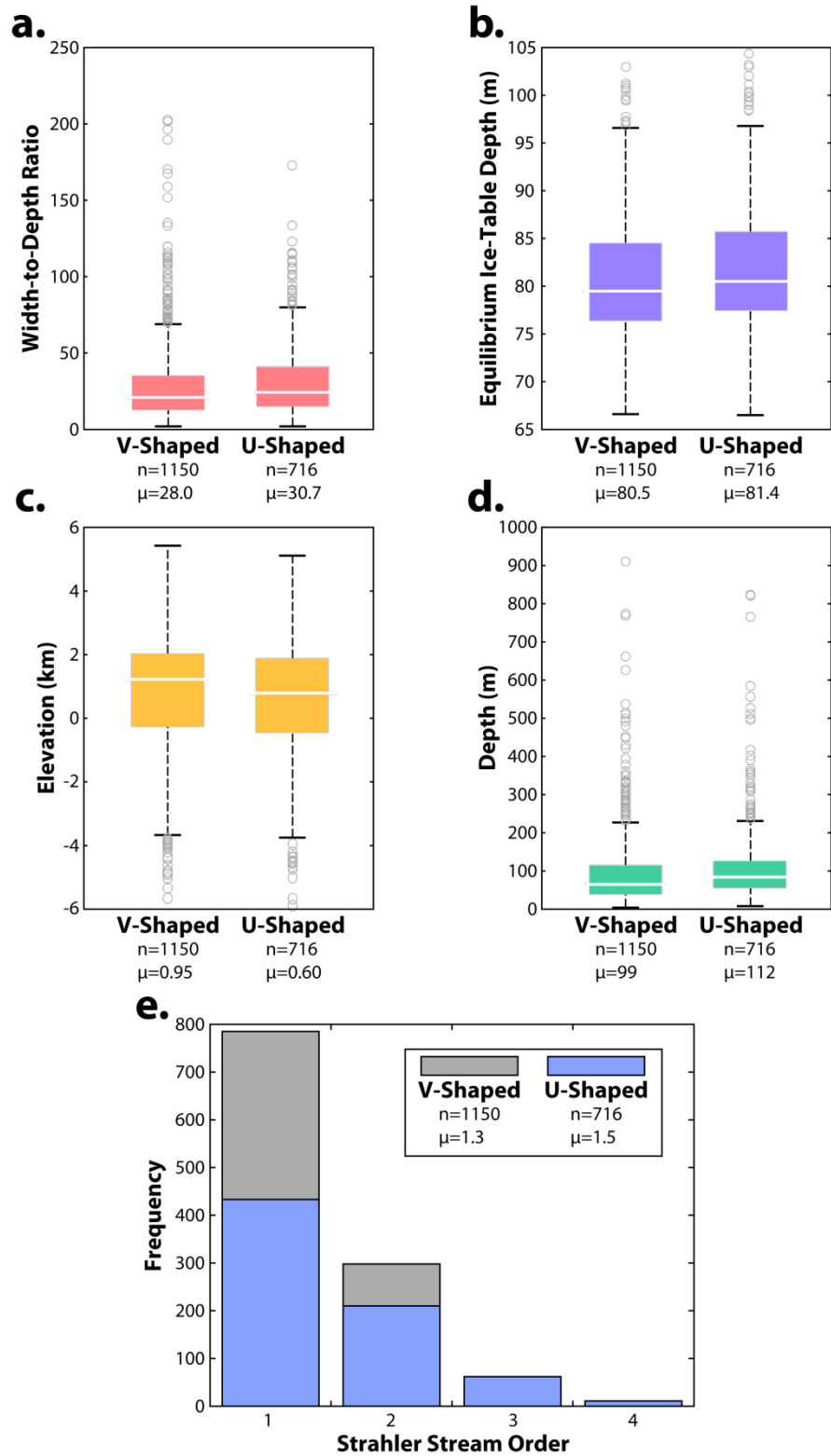


Figure 12. The distribution of U-shaped versus V-shaped valley network (a.) width-to-depth ratios, (b.) predicted ice-table depths, (c.) elevations, (d.) depths, and (e.) stream orders where n is the number of data points and μ is the mean value (data from Williams

and Phillips, 2001). Valley networks with U-shaped cross sections exhibit greater width-to-depth ratios (**a.**), but also occur in areas of slightly increased ice-table depths (**b.**) which conflicts with the predictions of the conceptual model (Figure 2). These characteristics may be due to the fact that U-shaped valleys tend to be more mature systems which occur at lower elevations (**c.**) and exhibit greater depths (**d.**) and stream orders (**e.**).

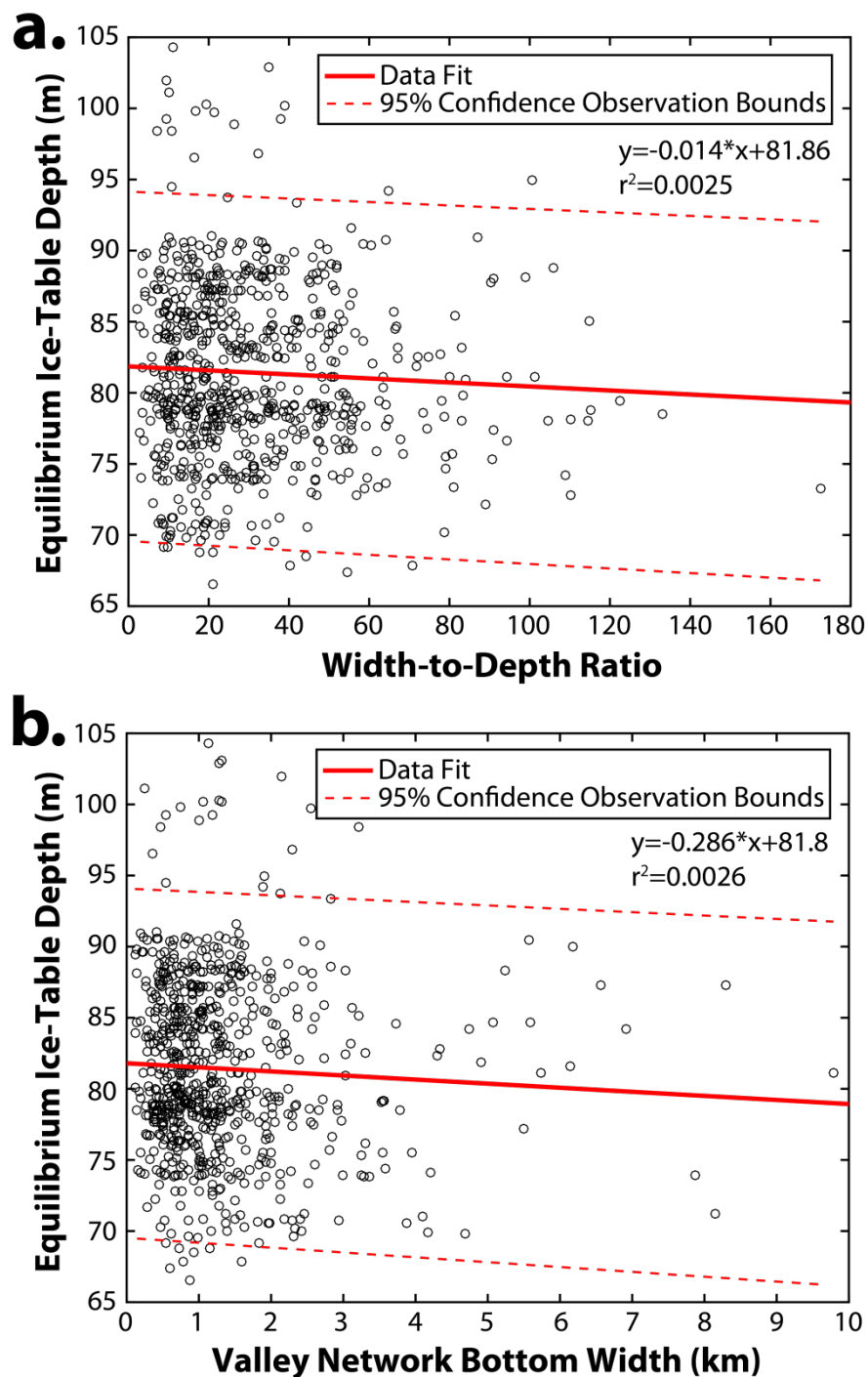


Figure 13. (a.) Measured (Williams and Phillips, 2001) valley network width-to-depth ratios versus predicted equilibrium ice-table depths for U-Shaped valley network systems. (b.) Measured (Williams and Phillips, 2001) U-Shaped valley network bottom width versus predicted equilibrium ice-table depths. In both panels the red line represents a linear regression fit and the dashed lines the 95% confidence observation bounds. No significant correlations are found between U-Shaped valley network width-to-depth ratios or bottom width and ice-table depth.

**Chapter 6: Evaluating the Hydrology of a Cold and Icy Early Mars: A
Synthesis**

James P. Cassanelli

Department of Earth, Environmental, and Planetary Sciences

Brown University

Providence, RI 02912, USA

1. Introduction

In light of recent global climate model predictions, this dissertation has explored the possible characteristics of a martian hydrologic system operating under cold and icy conditions. In this work, a new glaciovolcanic model has been proposed (Chapter 1) to account for the generation of liquid water and hydrologic features (fluvial features, in particular) under the predicted ambient cold and icy climate conditions. The volcanic and glacial processes involved in the proposed model are strictly endogenous to the martian system. Consequently, this proposed glaciovolcanic model for liquid water generation and hydrologic feature formation does not require the prevalence of a continuously warm and wet climate as has been inferred by previous studies. Thus, the glaciovolcanic work proposed (Chapter 1), developed (Chapters 1-3), and tested (Chapter 2-4) in this dissertation may provide a partial resolution of the inconsistency between the cold and icy climate conditions predicted by modeling efforts and the warm and wet climate conditions inferred from geomorphic studies as was introduced in the preface and discussed in preceding chapters. In order to test the validity of the proposed glaciovolcanic mechanism as an origin for martian hydrologic features, the model was applied and evaluated through case studies of the martian outflow channels, both those of the relatively older Noachian/Hesperian population (Chapter 2) and anomalously young Amazonian-aged population (Chapter 3), as well as a suite of features in the Tharsis volcanic province (Chapter 4). To further evaluate the potential existence of a cold and icy hydrologic system as explored in Chapters 1-4, the dissertation concluded with an assessment (Chapter 5) of a conceptual model for the formation of the martian valley networks under cold and icy conditions. This assessment was performed in order to

evaluate the consistency between the morphologic characteristics predicted by the model to result from valley network formation under cold and icy conditions and those observed from the global valley network morphology. This final chapter summarizes the major results from each of the preceding chapters, presents avenues for future work to address outstanding questions, and highlights the importance of these findings in the context of understanding the hydrologic and climatic history of Mars (and potentially other planetary systems) and the associated implications for habitability and the development of life.

2. Chapter 1: Lava heating and loading of ice sheets on early Mars: Predictions for meltwater generation, groundwater recharge, and resulting landforms

Global climate modeling studies of the early Mars climate (Forget et al., 2013; Wordsworth et al., 2013, 2015; Wordsworth, 2016; Palumbo et al., 2018) predict predominantly cold conditions under which liquid water is not stable at the surface of the planet. Given these conditions, how then, can the abundant presence of hydrologically-related features on the surface of Mars including the valley network systems (Howard, 2007; Fassett and Head, 2008a; Barnhart et al., 2009; Hynek et al., 2010; Hoke et al., 2011), open and closed-basin lakes (Carr, 1996, 2006; Cabrol and Grin, 1999; Fassett and Head, 2008b; Goudge et al., 2012), phyllosilicate-bearing units in Noachian-aged terrains (Bibring et al., 2006; Ehlmann et al., 2011), fluvially-degraded Noachian-aged impact craters (Craddock and Maxwell, 1993; Craddock et al., 1997; Craddock and Howard, 2002; Weiss and Head, 2015), and outflow channels (Baker and Milton, 1974; Carr, 1979, 1996, 2002; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b; Hanna and Phillips, 2005; Andrews-Hanna and Phillips, 2007; Cassanelli et al., 2015), be accounted for? This apparent paradox has been the subject of much interest, with

numerous mechanisms including volcanic greenhouse amplification (Halevy and Head, 2014), transient optimal insolation conditions (Wordsworth et al., 2015; Palumbo et al., 2018), and others proposed to cause punctuated heating and ice melting to generate these hydrologically-related features. These previously proposed mechanisms, however, require highly optimized, anomalous, or exogenous conditions and processes.

Therefore, in Chapter 1, the potential interaction of predicted early Mars regional ice sheets and spatially and temporally coincident extensive flood volcanism was explored (Cassanelli and Head, 2016) as a possible alternative mechanism to generate hydrologic features in an otherwise cold and icy environment. Using the Hesperia Planum region as a type area, we developed an ice sheet lava heating and loading model, quantitatively assessed the thermal and melting processes involved, and explored a broad parameter space to thoroughly constrain the mechanism and outline predictions for the formation of resulting geological features (Cassanelli and Head, 2016). The theoretical model developed in this chapter was then applied to a study area within the Hesperia Planum region to assess the consistency between derived model predictions and the observed geology. Ultimately, the ice sheet lava heating and loading process was found to be a viable mechanism to produce significant ice melting, infiltration, runoff, deformation, and to plausibly explain the characteristic geological features of the Hesperia Planum study area.

While the basic model concepts were developed and assessed through the groundwork performed in this chapter, many opportunities remain to improve and expand the theoretical treatment of the ice sheet lava heating and loading process as well as to

explore the geologic and hydrologic implications. These opportunities for future work are discussed below and in following sections.

2.1 What other glaciovolcanic processes may have occurred?

In the development and exploration of the ice sheet lava heating and loading model presented in this dissertation, we explored only a narrow range of possible lava-ice interaction geometries, specifically supraglacial lava emplacement. There exists, however, a multitude of potential geometries and processes by which extrusive volcanic processes could interact with surface ice deposits (Head and Wilson, 2007; Wilson and Head, 2009; Wilson et al., 2013; Edwards et al., 2014; Scanlon et al., 2014; Smellie and Edwards, 2016). Examples include: (1) the interaction of extrusive lava flows and surface ice deposits wherein the lava flow melts into and advances beneath the ice (a process observed to occur during the interaction of surficial ice deposits and lava flows on Earth; (Edwards et al., 2012, 2014, 2015; Smellie and Edwards, 2016), (2) snow deposition atop recently emplaced lava flows that have not completed cooling to ambient temperatures, and (3) deposition of heated tephra atop surface ice deposits (as previously considered in preliminary analyses; (Wilson and Head, 2009; Head et al., 2018). Modeling and evaluating this broader range of glaciovolcanic processes could help to better resolve the complete extent of the role glaciovolcanism has played in the hydrologic system of Mars and therefore inform our evolving understanding of the nature of the martian climate through time.

3. Chapter 2: Formation of outflow channels on Mars: Testing the origin of Reull Vallis in Hesperia Planum by large-scale lava-ice interactions and top-down melting

It is generally accepted that the enormous Hesperian and Amazonian-aged outflow channels of Mars were formed through diluvial activity (Baker and Milton, 1974; Carr,

1979, 1996, 2002; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b). However, the precise mechanisms responsible for the formation of these impressive erosional features remain unknown (Cassanelli and Head, 2018a, 2018b). The formation of the outflow channels has historically been attributed to catastrophic eruptions of groundwater from pressurized confined aquifers (Baker and Milton, 1974; Carr, 1979, 1996, 2002; Baker, 1982; Baker et al., 1992; Clifford, 1993; Head et al., 2003b; Harrison and Grimm, 2004, 2008; Andrews-Hanna and Phillips, 2007) resulting from excess hydraulic head generated by elevated recharge (Clifford, 1993; Harrison and Grimm, 2004; Russell and Head, 2007; Cassanelli et al., 2015) or from the downward propagation of a confining cryosphere layer from cooling of the climate (Carr, 1979, 1996, 2002; Hanna and Phillips, 2005; Wang et al., 2006). These models for outflow channel formation, however, have been challenged by difficulties in generating sufficient hydraulic head (Carr, 2002) and overpressure (Hanna and Phillips, 2005), requiring very high aquifer permeability (Head et al., 2003b; Manga, 2004; Harrison and Grimm, 2008), and by limited recharge sources (Russell and Head, 2007; Harrison and Grimm, 2009; Cassanelli et al., 2015).

Given the challenging conditions required by the canonical groundwater discharge models, in this chapter we assessed in detail the potential formation of outflow channels by the ice sheet lava heating and loading mechanism (Cassanelli and Head, 2018a). This assessment was performed through an applied case study of the Reull Vallis outflow channel system, selected for its close association with the effusive volcanic plains of the Hesperia Planum region (Crown et al., 1992, 2007; Crown and Greeley, 1993; Ivanov et al., 2005; Kostama et al., 2007; Gregg and Crown, 2009; Williams et al., 2009; Mest et

al., 2012; Tanaka et al., 2014). In order to perform this analysis, we assessed the geomorphology of the Reull Vallis system to outline required formation criteria which were then compared against model predictions generated utilizing regional glacial and lava emplacement conditions. We concluded that the Reull Vallis outflow channel could plausibly have been formed through an ice sheet lava heating and loading scenario involving sequential phases of melting ice deposits, which re-accumulated under an ambient “icy highlands” climate (Forget et al., 2013b; Wordsworth et al., 2013, 2015; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth, 2016) to undergo continual lava-ice interactions. We found this mechanism to offer several advantages over previous formation models, primarily in terms of its ability to supply the requisite volumes of water and to produce morphologic characteristics broadly consistent with those observed in the outflow channel system (Cassanelli and Head, 2018a).

While we concluded that the ice sheet lava heating and loading model could offer an advantageous explanation for the origin of the Reull Vallis outflow channel, there are several residual difficulties associated with the formation of the channel system by this mechanism which represent areas of required future research.

3.1 Insufficient meltwater generation area

The volume of meltwater produced by the ice sheet lava heating and loading mechanism is controlled in part by the total area of supraglacial lava emplacement and ice melting (Cassanelli and Head, 2016, 2018a). The areal extent of melting produced by this mechanism is predicted (Cassanelli and Head, 2016, 2018a) to be distinguishable by the extent of apparent subsidence and collapse features (due to the subsidence predicted to result from melting and removal of ice beneath the superposed lava). Through analysis

of the Reull Vallis system we found that, for plausible thicknesses of surface ice deposits, the total possible melting area at the head of the channel system (delineated by the extent of subsided and collapsed terrain) was insufficient to generate the volume of water required to form the outflow channel. To address this difficulty, we proposed an ice sheet lava heating and loading formation scenario for the Reull Vallis channel involving numerous phases of melting of thinner ice deposits facilitated by the re-accumulation of surface ice after melting and removal due to ambient “icy highlands” climate (Forget et al., 2013b; Wordsworth et al., 2013, 2015; Head and Marchant, 2014; Fastook and Head, 2015; Wordsworth, 2016) conditions. This proposed scenario requires episodes of lava flow emplacement with interspersed surface ice deposition occurring rapidly enough to accumulate deposits between lava emplacement events. To better evaluate the plausibility of this model, future work is required to: (1) constrain the total number and timing of lava flow emplacement events involved in the accumulation of the plains at the head of the Reull Vallis system, and (2) determine the timeframe required for the re-accumulation of surface ice deposits. The first suggested avenue of future work could be pursued through the acquisition and analysis of high resolution imagery at the Reull Vallis channel head to examine subsurface exposures produced by collapse in the area to identify lava flow layers to constrain their thickness and number. These estimates, in conjunction with estimates of the total accumulated regional lava thickness and emplacement timeframe could provide constraints on the number and timing of individual lava flow accumulation events within the vicinity of the Reull Vallis channel head. The second avenue of future work could potentially be addressed through meso-scale climate modeling (integrated into a global climate model) at the head of the Reull Vallis system to evaluate the

timescales required for the accumulation of surface ice deposits.

3.2 Ice sheet lava heating and loading meltwater fluxes

Thermal modeling analysis results (Cassanelli and Head, 2016, 2018a) indicate that the melt fluxes generated by the ice sheet lava ice heating and loading process alone are likely insufficient to directly account for the formation of the Reull Vallis outflow channel. In order for adequate meltwater discharges to be supplied by ice sheet lava heating and loading we proposed the potential involvement of meltwater impoundment and release processes. As evidenced by floods generated in terrestrial glacial environments (Bretz et al., 1956; Baker, 1973; Baker and Milton, 1974; Baker and Bunker, 1985; Costa and Schuster, 1988; Smellie, 2001; Benito and O'Connor, 2003; O'Connor and Costa, 2004; Das et al., 2008; Burr et al., 2009; Liang et al., 2012; Doyle et al., 2013; Tedesco et al., 2013), there exist a number of mechanisms which could have produced meltwater impoundment and release at elevated discharge levels. However, further work is required to more definitively identify the most plausible of these possible meltwater impoundment and release mechanisms (Cassanelli and Head, 2018a) and to determine if adequate discharge values could be reached under the conditions specific to the Reull Vallis system.

3.3 Outflow channel discharge

In this work (Cassanelli and Head, 2018a), we constrained the paleo-discharge of the Reull Vallis outflow channel through minimum and maximum end-member estimates based on minimum grain transport and bankfull flow assumptions (Wilson et al., 2004; Lamb et al., 2008; Lamb and Fongstad, 2010). These end-member estimates yield a wide range of possible paleo-discharges spanning several orders of magnitude, thereby

introducing large uncertainties into the assessment of channel formation mechanisms. Therefore, further analyses to better constrain the paleo-discharge of the Reull Vallis system would greatly improve the accuracy with which genetic mechanisms for the Reull Vallis system could be evaluated. In addition, more realistic estimates of the paleo-discharge would likely relax the discharge requirements for channel formation, as the bankfull flow assumption, adopted out of necessity, yields a strict maximum discharge (Cassanelli and Head, 2018a). Consequently, reductions in the estimated discharge necessary for channel formation could potentially allow for processes such as the ice sheet lava heating and loading mechanism to directly account for channel formation without requiring the operation of additional mechanisms to augment the supplied discharges.

4. Chapter 3: Large-scale lava-ice interactions on Mars: Investigating its role during Late Amazonian Central Elysium Planitia volcanism and the formation of Athabasca Valles

In Chapter 2, we examined the viability of the ice sheet lava heating and loading mechanism to account for the formation of Reull Vallis (Cassanelli and Head, 2018a), a typical member of the Noachian/Hesperian outflow channel population. In Chapter 3, we extended the model to investigate the outflow channels of Central Elysium Planitia (Cassanelli and Head, 2018b), an anomalously young population of outflow channels interpreted to have been active late into the Amazonian period. The anomalously young geologic age of the Elysium Planitia outflow channels (Plescia, 1990, 1993; Burr et al., 2002a, 2002b; Voigt and Hamilton, 2018) results in several demanding requirements for the canonical groundwater outburst formation mechanism. (1) A liquid groundwater system must have been preserved at great depth beneath a very thick cryosphere to near the present day (against the relatively efficient action of vapor diffusive loss to the

cryosphere; (Clifford, 1991), and must have been supplied by a source of active groundwater recharge (most plausibly Elysium Mons, still an unlikely source of recharge late in the Amazonian period near the time of formation of the Elysium Planitia outflow channels; (Russell and Head, 2007; Cassanelli et al., 2015). (2) Dike emplacement or tectonic extension must have produced and maintained an open conduit across the entire ~5-10 km thick cryosphere (Clifford et al., 2010) to allow water outflow to the surface. (3) The permeability of the aquifer from which the floods emanated must have been unreasonably large, far exceeding those common of terrestrial aquifers at similar depths and scales (Head et al., 2003b). (4) The groundwater system must have been saturated to the base of the cryosphere to allow confinement and pressurization, requiring potentially significant volumes of liquid water (Clifford, 1993). This final requirement may be particularly challenging as recent work (Weiss and Head, 2017) assessing the global distribution of impact crater morphologies suggests the martian groundwater supply may have been exhausted by the start of the Amazonian period due to vapor diffusive loss to the growing ice-cemented cryosphere.

Due to the difficulties faced by the classical groundwater models, we revisited the origin of the Elysium Planitia outflow channels and explored large-scale lava-ice interactions as a possible alternative mechanism to contribute to, or account for, their formation (Cassanelli and Head, 2018b). Given the abundant evidence for lava (Vaucher et al., 2009), ice (both surface and ground) (Head et al., 2003a; Mellon et al., 2004; Kadish et al., 2010; Bramson et al., 2015), and past lava-ice interactions (Hamilton et al., 2010, 2011; Keszthelyi et al., 2010; Dundas and Keszthelyi, 2013) preserved in the Elysium Planitia region (or at comparable latitudes in the case of ice; Pedersen et al.,

2010; Pedersen and Head, 2011, 2010), this is a plausible mechanism to explain the formation of outflow channels in the region. This assessment was performed through an applied case study of Athabasca Valles, the most intensely studied and best characterized outflow channel system of the region. Geomorphic criteria derived from the review of the Athabasca Valles system were used to test against predictions of meltwater generation and geomorphology produced through the application of the previously developed large-scale lava-ice interaction model (Cassanelli and Head, 2016, 2018a).

In Chapter 3 we concluded that the formation of the Athabasca Valles outflow channel system (and possibly other outflow channels in the region) could plausibly have been the result of surficial lava-ice interactions, but required meltwater confinement, accumulation, and release to supply the requisite discharges (Cassanelli and Head, 2018b). In preceding sections, we highlighted a number of avenues of future work required to better assess the involvement of surficial lava-ice interactions in the formation of outflow channels relating specifically to the supplied meltwater fluxes, channel discharge estimates, and meltwater confinement and release mechanisms. However, there are several remaining uncertainties regarding the formation of the Elysium Planitia outflow channels which could be addressed through future work not already discussed.

4.1 Relative age of volcanic plains and channel?

The ice sheet lava heating and loading model we tested to account for the origin of the Elysium Planitia outflow channels predicts a geologically contemporaneous age of channel formation and volcanic plains emplacement. The outflow channel systems of the Elysium Planitia region offer an ideal candidate to test this prediction as the systems are among the best preserved on Mars, and are inferred to have exceptionally young geologic

ages (~20 Ma) (as determined from embayment relationships with crater age-dated volcanic plains; Plescia, 1990, 1993; Burr et al., 2002a, 2002b; Voigt and Hamilton, 2018). Therefore, given the young, well-preserved state of the Elysium Planitia outflow channels, it may be possible to date any exposed erosive surfaces in the distal reaches of the channels not resurfaced by volcanic activity. Comparable age estimates for any formation related channel erosive surfaces and the volcanic plains which embay the system would support, though not conclusively, the origin of the outflow channels by ice sheet lava heating and loading while grossly different age estimates would suggest otherwise.

4.2 Where are the evaporites?

Formation of the Elysium Planitia outflow channels through the canonical groundwater effusion model would be anticipated to involve the release of groundwater confined within the martian crust for prolonged periods of time (Clifford, 1993). Consequently, prior to release to the surface, the groundwater would be expected to have equilibrated to subsurface conditions at elevated pressures (necessary to form the outflow channels by the outburst model) and thus to potentially contain a large load of dissolved solids (Clifford et al., 2010). Release of this groundwater to the surface and subsequent evaporation within terminal lows could therefore be expected to deposit spectroscopically detectable (Osterloo et al., 2008, 2010) quantities of evaporites (Tosca and McLennan, 2006). Additionally, the exceptionally young estimated age of the Elysium Planitia outflow channels would favor the preservation of any evaporite deposits produced in terminal lows/basins at the distal end of the channels. In contrast, outflow channel formation by surficial lava-ice interactions would involve erosion by relatively fresh ice-

derived meltwater not expected to contain a large load of dissolved solids. Therefore, identification of any evaporite deposits through focused spectroscopic observations at the distal end of the channels, or by the examination of previously generated global salt detection catalogs (Osterloo et al., 2008, 2010), would support a groundwater origin of the Elysium Planitia outflow channel systems.

5. Chapter 4: Glaciovolcanism in the Tharsis volcanic province on Mars: Implications for regional geology and hydrology

The Tharsis region of Mars is the most prominent and active center of volcanism on the planet, with a history of volcanic activity spanning from the Noachian to the very recent Late Amazonian period (Solomon and Head, 1982; Tanaka et al., 1988; Phillips et al., 2001; Werner, 2009; Carr and Head, 2010; Mangold et al., 2010; Richardson et al., 2017). Due to its significant elevations and the influence of adiabatic cooling, the Tharsis province is also a region predicted (Forget et al., 2013b; Wordsworth et al., 2013, 2015; Wordsworth, 2016) to accumulate regional ice sheets under both “cold and icy” (global mean annual surface temperature ~ 225 K; Wordsworth et al., 2015) and “warm and wet” (global mean annual surface temperature > 273 K; Wordsworth et al., 2015) end-member climate scenarios. Beyond the early history of Mars, following the loss of the thicker atmosphere (Carr and Head, 2010; Lammer et al., 2013; Jakosky et al., 2017), abundant evidence (Head et al., 2003a; Shean et al., 2005; Forget et al., 2006; Fastook et al., 2008; Kadish et al., 2008; Head and Weiss, 2014) suggests regional accumulation of snow and ice continued throughout the Amazonian period within the Tharsis region. Therefore, the Tharsis region is predicted to be a site of coincident volcanic and glacial activity throughout the majority of the history of Mars, raising the possibility for widespread and sustained glaciovolcanism. In light of this prediction, an exploratory study was performed

in Chapter 4 to assess the potential role of glaciovolcanism in the geologic and hydrologic history of the Tharsis region (Cassanelli and Head, 2018c). This assessment was performed by first reviewing the history and characteristics of volcanism and glaciation in the Tharsis region, as well as previously documented evidence for past glaciovolcanic activity, to constrain relevant conditions and parameters. The potential role of glaciovolcanism was then evaluated in the formation of major tectonic and hydrologic features within the region by outlining geomorphic formation criteria and comparing against predictions from possible genetic surficial glaciovolcanic models informed by previous modeling work (Cassanelli and Head, 2016, 2018a). We concluded that surficial glaciovolcanism may plausibly have contributed to the formation of many of the tectonic and fluvial features in the Tharsis region, offering advantages over prior formation models, particularly for the large basin/chaos-sourced outflow channels concentrated in the area (Cassanelli and Head, 2018c). However, given the first-order, exploratory nature of this assessment of regional glaciovolcanism in the Tharsis province, many uncertainties remain that could potentially be addressed through additional future analyses which are discussed in the following sections.

5.1 Missing volcanic valley networks?

In this chapter, we examined (Cassanelli and Head, 2018c) the valley network systems incised into the flanks of the volcanic edifices of the Tharsis region, distinct in age and morphology from the broader population of valley networks spread across the highland terrains of Mars (Fassett and Head, 2006, 2007, 2008a; Hynek et al., 2010). We explored, through a case study of the Ceraunius Tholus edifice, an origin of these volcanic valley network systems by surficial glaciovolcanic interactions involving the

top-down melting of surface snow and ice by supra-ice lava flow emplacement. We concluded that this mechanism is better able to account for the formation of these particular valley network systems than the previously favored basal melting model (Fassett and Head, 2007) because: (1) the proposed mechanism provides greatly elevated meltwater production rates relative to basal melting, and (2) is not anticipated to produce morphologic signature of wet-based glaciation, consistent with the absence of these features on the volcanic flanks. However, interestingly, many of the largest volcanic edifices within the Tharsis region (Olympus Mons, Pavonis Mons, Arsia Mons, and Ascraeus Mons) exhibit the most sparse distributions of valley network systems, the lowest drainage densities, and the least developed systems (Hynek et al., 2010b). This observation contrasts with expectations of the proposed glaciovolcanic origin model given that these large volcanic edifices experienced the most intense, prolonged, and recent volcanic activity (Solomon and Head, 1982; Tanaka et al., 1988; Phillips et al., 2001; Werner, 2009; Carr and Head, 2010; Mangold et al., 2010; Richardson et al., 2017) and contain the strongest and most abundant evidence for the presence of glacial deposits (Shean et al., 2005; Forget et al., 2006; Fastook et al., 2008; Head and Weiss, 2014) and occurrence of glaciovolcanic activity (Kadish et al., 2008; Scanlon et al., 2014, 2015).

One potential explanation for the under-developed state of the valley network systems on the large volcanic edifices is the elevation of the summit regions. The great elevation of the summit regions of the large volcanic edifices may have prevented widespread interaction of heated lava flows and surface ice deposits through the generation and emplacement of lava flows at elevations exceeding those of major snow and ice accumulation (Fastook et al., 2008). An avenue of future work that may address the

enigmatic lack of developed valley network systems on the large volcanic edifices includes meso-scale climate modeling of the Tharsis region to evaluate the differences in the patterns of snow and ice deposition between the larger volcanic edifices and the smaller edifices. In particular, this climate modeling work could be used to evaluate the distance of ice deposition from the summit region as a function of edifice size and to assess potential correlations of this distance to the degree of valley network system development.

5.2 Origin of the basin-sourced Tharsis region outflow channels beyond Kasei?

In Chapter 4, we explored (Cassanelli and Head, 2018c) the viability of a surficial glaciovolcanic origin for the basin-sourced outflow channels of the Tharsis region through a focused case study of the Kasei Valles system. In addition to the Kasei Valles system, the Tharsis regions hosts a number of other basin-sourced outflow channels including the Maja, Shalbatana, Simud, and Tiu Valles systems. In this chapter, we suggested that these additional basin-sourced channels may have shared a similar surficial glaciovolcanic formation mechanism as Kasei Valles on the basis of their comparable morphology and location. However, the drainage basins feeding these other basin-sourced outflow channel systems are likely to differ significantly from that of Kasei Valles in their size, location, and proportion occupied by volcanic plains of the Tharsis province. As a result, the drainage basins of these other outflow channel systems may provide less potential for surficial glaciovolcanic interactions and meltwater production than that of Kasei Valles, thereby challenging a potential mutual glaciovolcanic origin mechanism (involving the melting of surface ice deposits in response to volcanic emplacement associated with the construction of the Tharsis rise). Therefore, additional

investigation is required to delineate and characterize the drainage basins of each of these additional outflow channel systems to account for local conditions and better evaluate potential formation of these systems by the surficial glaciovolcanic mechanism.

6. Chapter 5: Assessing the formation of valley networks on a cold early Mars: Predictions for erosion rates and channel morphology

Mars hosts a diverse array of valley networks, systems of linear-to-sinuuous depressions which are widely distributed across the surface and which exhibit branching patterns similar to the dendritic drainage patterns of terrestrial river systems (Carr and Clow, 1981; Baker et al., 1992; Carr, 1996, 2000, 2012; Hynek et al., 2010b). Characteristics of the valley networks are indicative of an origin by fluvial activity (Carr, 1995, 1996; Gulick, 2001; Fassett and Head, 2005; Howard et al., 2005), providing among the most compelling evidence for the past presence of liquid water at the surface of Mars. Stratigraphic and crater age dating techniques suggest that the formation of the valley networks occurred during the early geologic history of Mars (Fassett and Head, 2008a). However, whether these fluvial systems formed predominantly by rainfall in a relatively warm early Mars climate (Craddock and Howard, 2002), or by snowmelt and episodic rainfall in an ambient cold climate (Carr and Head, 2003; Palumbo et al., 2018), remains disputed. Understanding the formative environment of the valley networks will help distinguish between warm and cold end-member early Mars climate models (Wordsworth et al., 2015) which is important to understanding the full range of conditions and environments that might have supported life on Mars. To address the uncertainty regarding the nature of the environment in which the valley networks formed, we proposed and tested a conceptual model for valley network formation under the cold and icy conditions (Cassanelli and Head, 2018d) suggested by global climate models

(Forget et al., 2013b; Wordsworth et al., 2013, 2015) to have dominated early Mars history. Through a series of quantitative tests of the proposed conceptual model we outlined predictions for characteristic valley network morphologic features potentially produced as a result of formation under cold and icy conditions (Cassanelli and Head, 2018d). We then performed a global assessment of valley network morphometry to determine if the observed morphology was consistent with that predicted by the model for valley network formation in a cold and icy environment.

While results of the theoretical analysis (Cassanelli and Head, 2018d) indicated that the morphology of the valley networks could plausibly have been influenced by cold and icy conditions, global examination of the valley network morphometry produced inconclusive results. The indeterminate nature of the global morphometric analysis was attributed to a number of complicating factors and simplifying assumptions that require further investigation as discussed in the following sections.

6.1 Heterogeneous substrate properties

In the theoretical analysis of the conceptual model proposed in Chapter 5, we assumed constant and globally homogeneous substrate properties including the geothermal heat flux and thermal conductivity. These assumptions represent clear oversimplifications of the subsurface properties which are certain to vary significantly across the surface of Mars. In addition, this analysis was performed with the use of global climate model data outputs assuming a typical Late Noachian early Mars climate scenario characterized by a 1 bar CO₂ atmosphere and 45° planetary obliquity (Forget et al., 2013b; Wordsworth et al., 2013, 2015; Palumbo et al., 2018). While both of these climate model parameters are expected to vary considerably with time and for different

climate scenarios (Laskar et al., 2004; Forget et al., 2013b; Wordsworth et al., 2013), in this assessment we considered only the base-line scenario. Therefore, given these simplified model inputs, there is room for significant improvements to the theoretical treatment presented in Chapter 5 through the use of more robust data inputs generated by further research. Variations in the geothermal heat flux and substrate thermal conductivity across the surface of Mars could be taken into account, respectively, through the implementation of: (1) global geothermal heat flux distributions derived from geodynamic modeling (Plesa et al., 2016), and (2) a global estimate of the substrate thermal conductivity distribution constrained with the use of in situ calibrated global thermal inertia and albedo remote sensing data sets (Williams et al., 2013; Jones et al., 2014; Harrison et al., 2015; Rogers et al., 2018). Potential variations in the prevailing climate parameters could be accounted for through the exploration of a wider climate scenario parameter space to consider variations in the planetary obliquity, atmospheric pressure, and other properties (Forget et al., 2013b; Wordsworth et al., 2013, 2015; Palumbo et al., 2018).

6.2 Valley network morphologic/morphometric considerations

In Chapter 5, we utilized a global morphometric data set (Williams and Phillips, 2001) in order to assess the consistency of the geomorphology of the martian valley networks with the predictions of the conceptual and quantitative models for valley network formation under cold and icy conditions. To perform this assessment, we specifically examined the valley network width-to-depth ratios, cross-sectional morphology (V-shaped versus U-shaped), and valley bottom width. Results of our analysis (Cassanelli and Head, 2018d) revealed poor correlations between these valley

network morphological parameters and trends predicted from the conceptual and quantitative modeling. However, these morphological parameters may have been subject to alteration by a number of factors and processes not accounted for in our analysis that would require consideration in order to more rigorously evaluate the predictions of the conceptual and quantitative models developed in this chapter. Factors and processes that may have influenced the examined valley network morphometric parameters which require further investigation include: (1) Substrate heterogeneity: In our assessment, we assumed constant, globally homogeneous substrate properties. Variations in the composition and physical state (*e.g.* porosity) of the substrate across Mars, however, would likely have caused variations in the rates and patterns of fluvial erosion and, by consequence, valley network morphometry. To better account for the variable nature of these substrate properties, global thermal inertia, albedo, and infrared spectroscopy data (Williams et al., 2013; Jones et al., 2014; Harrison et al., 2015; Rogers et al., 2018) could be utilized to broadly constrain the composition and physical state of the near-surface substrate materials across the martian surface . (2) Valley preservation state: The morphometric data utilized to perform the morphological assessment in Chapter 5 are necessarily derived from measurements of the preserved, relict fluvial valley systems that constitute the valley network population. As a result, the available morphometric measurements are not perfectly representative of the original fluvial channels (see Carr, 1996, p. 47-100 for discussion of the importance of this distinction) as they incorporate the influence of other processes and factors which have modified the post-formation valley profiles. Processes and factors that may have resulted in modifications to the post-formation valley profiles include mass wasting processes, aeolian erosion and infilling,

modification by glacial processes (*i.e.* lineated valley fill; Head et al., 2006a, 2006b, 2010), the strength/cohesion of the substrate materials in which the valley formed (Sklar and Dietrich, 2001), and the angle of repose of the substrate material. As a result, in order to retrieve the best representation of the morphological signature imprinted by the prevailing climate conditions at the time of valley formation requires further analysis to (a) account for and eliminate the influence of the myriad of post-formation processes and complicating geological factors, (b) identify and assess only the least modified valley systems, or (c) assess only those valley systems likely to have experienced similar modification histories. (3) Sampling bias: The global data set (Williams and Phillips, 2001) used to perform the morphometric analysis contained ~2,000 measurements of valley networks, with measurement points dictated by locations where MOLA crossings of valley network systems occurred at the time of the study. Therefore, an updated, more complete and consistent sampling of valley network morphometry might yield an improved correlation between ice-table depth and width-to-depth ratios. (4) Valley network measurement position: Here we performed an indiscriminate evaluation of a complete global data set (Williams and Phillips, 2001) of valley network morphometric measurements. However, individual valley network systems are likely to experience variations in substrate and flow conditions between and along the tributary and trunk segments. These variations, and their potential influence on valley morphometric parameters, may also be compounded further when examining different valley network systems. Examples of possible variations include changes in substrate erosive resistance (*e.g.* loose regolith versus consolidated material) and valley floor slope (Sklar and Dietrich, 1998, 2001). Therefore, to isolate these influences, future analyses should be

conducted on comparable sections of the same or different valley network systems which may have experienced comparable substrate and flow conditions and on individual valley network systems as a function of distance from head to terminus.

6.3 Earth analog?

A final avenue of future research which may aid in understanding and recognizing the potential influence of cold and icy conditions on the formation and morphology of hydrologic features on Mars is the examination of Earth analogs. Of the many terrestrial environments hosting hydrologic features, the conditions present within the McMurdo Dry Valleys of Antarctica represent the closest known example to those predicted to characterize early Mars during the time of hydrologic feature formation (Head and Marchant, 2014). As a result, the McMurdo Dry Valleys have been utilized as an Earth analog site by a number of previous studies (Dickson et al., 2013; Head and Marchant, 2014) aiming to develop insight into the climatic conditions of early Mars. Despite classification as a hypothermal, hyper-arid, polar desert, with air temperatures dropping to as low as -50°C during the austral winter (Dickson et al., 2013), the McMurdo Dry Valleys contain a wide array of hydrologic features including gullies, fluvial systems, and lakes (the most striking example of which is Don Juan Pond exhibiting a salinity sufficient to keep the lake perennially ice-free) (Dickson et al., 2013; Head and Marchant, 2014). Of particular importance is the presence of a thick and regionally continuous permafrost (*i.e.* ice-cemented regolith) within the McMurdo Dry Valleys, a feature predicted (Cassanelli and Head, 2018d) to have had a primary influence on the formation of hydrologic features on early Mars. Given the similarities in environmental conditions between the McMurdo Dry Valleys and those predicted to characterize early Mars,

assessing the evolution of the suite of active hydrologic features of interest within the McMurdo Dry Valleys could provide valuable analog insight into the operation of hydrologic processes on Mars.

7. Summary and Broader Impacts

In this dissertation, we explored the nature of the martian hydrological cycle existing under cold and icy climate conditions which models predict to have prevailed for the majority of the history of the planet. In Chapter 1 we proposed and developed a new glaciovolcanic model as a mechanism to generate liquid water at the surface of Mars in an ambient cold and icy climate, thereby providing a possible explanation for the origin of the variety of hydrologic features preserved across the surface of the planet. In Chapters 2 through 4, we further refined the glaciovolcanic model proposed in Chapter 1 and evaluated its viability as a genetic mechanism for several martian hydrologic features with a focus on the valley networks and outflow channels. Through the investigations of Chapters 1 through 4, we concluded that large-scale surficial glaciovolcanic interactions could plausibly have contributed to the formation of the valley network and outflow channel systems. Future work, however, is required to evaluate a more complete range of possible glaciovolcanic processes that may have taken place, mechanisms that could augment the relatively low meltwater fluxes produced by glaciovolcanic processes, and observational evidence potentially supportive of a glaciovolcanic origin for martian hydrological features (*e.g.* an absence of evaporite deposits at fluvial feature termini). These future analyses will involve additional modeling and quantitative analyses as well as geomorphic and spectroscopic observational assessments.

In Chapter 5, we proposed a conceptual model for valley network formation and incision under cold and icy climate conditions. Through quantitative evaluations of the

proposed conceptual model, we outlined predictions for characteristic valley network morphologic signatures that could potentially have been produced as a result of formation in a cold and icy environment. We then performed a global assessment of valley network morphometry to determine if the observed morphology was consistent with that predicted by the proposed model for valley network formation in a cold and icy environment. Results of the quantitative analysis of the conceptual model presented in Chapter 5 indicated that the morphology of the valley networks could plausibly have been influenced by cold and icy climate conditions. However, the global examination of valley network morphometry to assess consistency with the model predictions produced generally inconclusive results. The indeterminate nature of the morphometric analysis was attributed to a number of factors and simplifying assumptions built into the quantitative treatment that require further investigation to generate a more definitive assessment. The required further investigation will involve accounting for variable substrate properties (such as geothermal heat flux and thermal conductivity), a wider range of possible climate scenarios, the preservation state of the valley networks under examination, and culling the morphometric data to analyze valley network systems with comparable histories. This future work can be addressed through the utilization of global substrate property distributions derived from models and remote sensing measurements, additional global climate model results, and a more discriminate analysis of the valley network morphometric data.

The viability of the glaciovolcanic model as a mechanism to generate liquid water under ambient cold and icy conditions has potentially profound implications for the climatic and hydrologic history of Mars. Important among these implications is the

potential of this process to have contributed to the formation of the martian hydrologic features including valley networks, outflow channels, and open and closed-basin lakes as discussed in Chapters 1-4. The contribution of the glaciovolcanic ice sheet lava heating and loading process to the formation of these hydrologic features, in part, negates the inferred requirement of a warm and wet early climate, vertically integrated hydrologic system (*i.e.* communication of the surface and subsurface hydrologic systems), anomalous heating conditions, and difficult subsurface groundwater system conditions (*i.e.* exceptionally high permeabilities). Therefore, this model can help reconcile the predictions of a cold and icy early Mars climate and the presence of hydrologic features by offering a viable formation mechanism able to operate under the predicted conditions. Furthermore, geomorphic evidence supporting the operation of the glaciovolcanic ice sheet lava heating and loading process can be used as an indication of the presence, distribution, and timing of surface ice deposition on Mars and therefore of cold and icy climatic conditions. Beyond Mars, it is physically plausible for the glaciovolcanic ice sheet lava heating and loading process to have occurred on other planetary bodies hosting regional surface ice deposits and generating effusive volcanism. Most notable among the other planetary bodies on which these processes may have occurred is Earth, where evidence of the ice sheet lava heating and loading process might be expressed in terrestrial large igneous provinces that have been emplaced at high latitudes. Finally, while the results of the analysis of Chapter 5 were inconclusive, suggested future work holds the potential to demonstrate a signature of cold and icy formation conditions preserved in the morphology of the valley networks. The clear implication of this potential future demonstration is the formation of valley networks by melting of surface

snow/ice or punctuated rainfall produced by transient heating events, therefore indicating a predominantly cold and icy climate during early Mars history in agreement with climate model predictions. Ultimately, the implications of the findings of this dissertation, and those possible through the suggested avenues of future investigation, relate to the presence and nature of habitable environments on Mars and the potential for the development and support of life. If indeed the early history of Mars was dominated by a cold and icy climate and hydrologic system as explored in this dissertation, this suggests a more restricted range of possible habitable environments, most likely within the subsurface. Given the emerging directives of the NASA Mars exploration program to “explore habitability” and “seek signs of life” this is an important practical determination to help guide future exploration efforts.

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