

**Clues to Martian Paleoclimate:
Ice Reservoirs, Sediment Transport, and Depositional
Environments**

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Submitted in partial fulfillment of the requirements for the degree of Doctor of
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*We have an infinite amount to learn both
from nature and from each other. – John Glenn*

In January of 1990, I was five and a half years old – staying up far past my bedtime, transfixed by grainy black and white video beamed down from the Space Shuttle *Columbia*. STS-32 was a mission to recover NASA’s Long Duration Exposure Facility (LDEF), a task delayed by more than four years following the loss of *Challenger*. I sat transfixed watching the astronauts’ plodding approach to LDEF, their careful survey of its many facets, and finally their gentle grappling, and stowage of the 21,400 pound experiment into the cargo bay. I recorded the whole thing on VHS tape and couldn’t wait for the launch of the Hubble Space Telescope in the spring. I’ve always been interested in science, space, and technology. At some point I started looking at the rocks beneath my feet as well...

Thank you to Nick Christie-Blick, my undergraduate advisor who introduced me to the Basin-and-Range. The faculty of the department of Earth and Environmental Sciences faculty at Columbia enriched my undergraduate experience with many great courses and fieldtrips. I owe special thanks to Bill Ryan who advised me in my first research project. I

learned so much about the process of science that summer working with old seismic data.

Thank you to the graduate students who have come before me. Sometimes it may feel like a solitary endeavor, but the camaraderie and support of my fellow students cannot be overstated. Specific stories and names are probably best left to oral histories – thank you all for the good times. Special thanks are owed to Joseph Levy and Caleb Fassett who have been both helpful and inspirational along the way. I am indebted to all the students with whom I’ve shared my time at Brown. I hope in some small measure, I’ve been able to pass on to those coming along, all I’ve learned from those who went ahead of me.

Thank you to Jim Head, my academic advisor. It has been a great privilege and opportunity to be part of your research group. The experience, data, and opportunities that you have shared have been tremendous. I’ve also learned a lot from you about science and life, -- and about clarity in scientific writing, even if I still insist on using “evidence” in its verb form now and then... Thank you to my committee. Your courses, questions, guidance, and perspectives have helped me greatly along the way. Thank you: Carle Pieters, Tim Herbert, Jim Russell, and Dave Marchant. I also owe special thanks to Jay Dickson, Nancy Christie, and Lucia Tavares who never failed to brighten my days in the Lincoln Field Building.

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*Experience: that most brutal of teachers.
But you learn. My God do you learn. – C.S. Lewis*

PREFACE

Good hockey players are born in the first quarter of the year. Star baseball players have August birthdays. -- Thanks to Malcolm Gladwell for being a relentless proponent of the idea that *timing really matters*.

I've been privileged to be working during a golden age of Mars exploration. The first high-resolution images of the Martian surface were acquired while I was in high school -- just in time to capture my interest as an undergraduate student for a summer internship at the US Geological Survey's astrogeology branch in Flagstaff, Arizona. From there, I entered graduate school in the fall of 2006 just as a torrent of data began arriving from *Mars Reconnaissance Orbiter*. If I have seen farther than others, it is because I have stood on the shoulders of scientific giants and have had the privilege to peer down at Mars through the largest telescope ever sent beyond Earth's orbit.

Sedimentary systems provide a foundational means of understanding the workings of the natural world. Sedimentary deposits serve as a lens through which geoscientists are able to study topics ranging from tectonics to

climate variability. These deposits are a record of geologic time; they are a history that contains markers of significant events as well as a measure of prevailing conditions. Through their erosional modification and alteration, sedimentary deposits even record subsequent conditions well after their deposition. Terrestrially, the study of sedimentary systems has been integral to advances in fields as disparate as continental tectonics, glaciology, paleoclimatology, as well as petroleum geology. Now these research approaches are applicable to Mars.

By focusing specifically on depositional environments and processes, I am able to ascertain bounding characteristics about the prevailing climate system, volatile reservoirs, and aqueous transport mechanisms during specific portions of the Noachian and Amazonian that are germane to major uncertainties regarding climate variability. More specifically, a tenuous scientific consensus now supports a so-called “warm-and-wet” view of Noachian Mars; however, a significant debate exists over the actual persistence of such conditions – an issue the sedimentary record is uniquely suited to address. Similarly, Amazonian latitude-dependent landforms such as gullies, layered mantling deposits, and valley glacial features suggest large fluxes between discrete ice reservoirs. Remaining fundamental questions about the transport mechanisms, periodicity, timing, and extent of these fluxes are tractable through detailed stratigraphic analysis. A subset of

such questions is addressed here with techniques that are extensible to other geological problems.

Chapter 1 considers the climatic and sediment transport history of Mars' Noachian through an analysis of the Jezero crater paleolacustrine system. The Noachian has often been thought of as at least partially “warm-and-wet.” A fundamental question, also of interest in astrobiology, concerns the duration and stability of these conditions. Were temperate episodes violent, punctuated occurrences of hydrologic activity separated by arid, impact-dominated interludes? Or, did Mars' fluvial and lacustrine geomorphologies form under a relatively stable climate that was conducive to overland flows of biologically meaningful duration? Did the final termination of these conditions come gradually or was it abrupt?

The Jezero paleolacustrine basin system and delta assemblage requires significant hydrologic-climatic stability during the period of fluvial-lacustrine activity and constrains potential climate episodicity during the Noachian by requiring a minimum duration of clement conditions. Residual accommodation space in the basin indicates that overland flow and sediment transport were fundamentally limited. The preserved sedimentary deposits suggest a relatively abrupt termination of the hydrologically active period. Therefore, while the Noachian experienced clement and stable conditions consistent with a warm-and-wet early Mars during construction of the Jezero

delta complex, such conditions are not required in sufficient duration to be considered representative of Mars early history in general.

Chapter 2 steps forward dramatically in geologic time to the Late Amazonian, long after the permanent transition to hyperarid, hypothermal, polar-desert-like conditions. This chapter presents a novel technique (employing secondary crater clusters as a chronostratigraphic marker) for dating the development of Mars gully deposits. The origin of gullies on Mars has been controversial (e.g., formative hypotheses include catastrophic groundwater release, debris flows, dry granular flows, and meltwater from surface ice and snow). The age constraint presented in this chapter conclusively links gully activity to recent geologic history (≤ 1.25 Ma). The chronological marker developed in this chapter is coincident with the emplacement of dust-ice mantling deposits interpreted to represent recent ice ages on Mars. The relationship between these deposits and the formation of gullies is addressed further in Chapter 3. This association, together with multiple episodes of depositional fan formation (mapped in this chapter), favors an origin for gullies from top-down melting of snow and ice during multiple favorable spin-axis and orbital variations.

In Chapter 3 the genetic relationship between ice-rich mantling deposits and gully formation is explored in significantly more detail. While co-observations of gullies and latitude-dependent mantling have been made previously, in this chapter new observations of small surficial gullies that

lack alcoves are described for the first time and placed in chronostratigraphic context. These features strongly indicate that degradation of the mantling deposits (ablation of ice content) is responsible for generating aqueous phases (meltwater) to form these gullies. The lack of alcoves suggests that local accumulation zones are not required for their formation. Because the mantling unit is linked to glacial episodes of a recent ice age pre-dating the stratigraphic marker, we suggest that this type of gully formation has waned in the current interglacial conditions that have prevailed following a damping of Mars obliquity, ~400 ka.

Chapter 4 completes the gully analyses with an investigation of potentially anomalous gullies within Gasa crater. The geologic setting reveals that the recent Gasa impact event occurred into an ice-rich crater interior. By providing meltwater and an erodible substrate, this unique impact environment is responsible for the formation of Gasa's large gullies. Evidence of melting, flow, and ponding of Gasa substrate and ejecta is observed on the wall and floor of the host crater. Neighboring craters provide additional geological evidence of contemporaneous accumulations of glacial ice. Because the Gasa host crater contains latitude-dependent mantling deposits in conjunction with canonical gully features as well alcove-less gullies, this locale provides a setting where the duration of recent gully formation can be explored and placed in the context of recent ice ages.

In Chapter 5 attention turns to geological evidence of Mars recent ice ages and the fluxes of ice associated with these events. Global climate models predict that ice will be deposited in tropical regions during obliquity excursions from the current mean obliquity of $\sim 25^\circ$ to $\sim 35^\circ$, but no geological evidence for such deposits had previously been found. Here, the presence of a series of very small (decameter-scale) pedestal craters in the tropics of Mars (the Daedalia Planum-Tharsis region) that are superposed on an impact crater dated to ~ 12.5 million years ago are reported for the first time. The characteristics, abundance, and distribution of these small pedestal craters provide geological evidence that meters-thick ice accumulations existed in the tropical Tharsis region of Mars in the last few million years when mean obliquity was $\sim 35^\circ$ (~ 5 -15 Ma) before it transitioned to a mean of $\sim 25^\circ$. Similar small pedestal crater examples are superposed on the older Amazonian Arsia Mons tropical mountain glacier deposit, suggesting that ice can accumulate in these tropical regions without initiating large-scale glacial conditions. These results provide the first support for predictions of general circulation models that ice migrates to the equatorial regions during periods of moderate obliquity and then serves as a source for mid-latitude ice-age deposits from there.

Finally, Chapter 6 utilizes rayed and young craters to explore the latitude-dependence of ice-age related mantling deposits. Rayed craters are the youngest impact craters on a planetary surface. Understanding the age

and distribution of these craters helps elucidate surface processes and resurfacing histories that provide important constraints on Late Amazonian climate variations. By counting more than 60,000 superposed craters, this study also provides important new data regarding crater-size frequency distributions of the smallest impact craters. These results confirm the applicability of the isochron system using small craters. This chapter examines a collection of km-scale young craters and through detailed counts of craters superposed on their near-rim ejecta deposits calculates crater retention ages. The distribution and ages of these young craters supports geologically-recent deposition of ice-rich mantling deposits in the mid-latitudes poleward of approximately 30 degrees during the last few million years.

The final chapter integrates results described in the previous research chapters to review implications for Mars broader climate history, particularly the Late Amazonian, in the form of outstanding questions. Special focus is paid to outstanding research questions arising from the geological evidence for mid-latitude glaciations and the genetic relationship between ice-rich mantling deposits during recent ice ages and gully processes. The chapter ends with a discussion of exploration strategy and the appropriateness of embarking on a Mars sample return mission at this time.

TABLE OF CONTENTS

SECTION	PAGE
Signature Page	iii
Curriculum Vitae	iv
Acknowledgements	x
Preface	xiii
Chapter 1: Implications for Noachian Climate from an Overfilled Lacustrine System and Progradational Delta in Jezero Crater, Mars.	
Abstract	1-1
Introduction	1-4
Classification of Lakes	1-8
Lakes on Mars	1-11
Jezero Lacustrine System	1-13
Deltaic Deposits	1-18
Delta Plain Sedimentary Features	1-23
Discussion	1-25
An Overfilled Lacustrine System	1-27
Sedimentation Rates	1-30
Comparison to Modeling Results	1-34
Conclusions	1-35
Acknowledgments	1-38
References	1-38
Figures	1-70

Chapter 2: Unique chronostratigraphic marker in depositional fan stratigraphy on Mars: Evidence for ca. 1.25 Ma gully activity and surficial meltwater origin.

Abstract	2-1
Introduction	2-2
Gully Fan Stratigraphy	2-4
Nearby Rayed Crater Source of Secondary Craters	2-5
Discussion and Conclusions	2-8
Acknowledgements	2-10
References	2-11
Figures	2-19

Chapter 3: Keys to Gully Formation Processes on Mars: Relation to Climate Cycles and Sources of Meltwater.

Abstract	3-1
Introduction	3-2
Dating Gully Formation	3-3
Gullies in a Stratigraphic Context	3-5
Degradation of the Latitude-dependent Mantle	3-7
Alcove-less Gullies Suggest a Candidate Source of Meltwater	3-9
Activity During the Current “Interglacial” Period	3-11
Conclusions and Implications	3-13
Acknowledgements	3-14
References	3-14
Figures	3-22

Chapter 4: Gasa Impact Crater, Mars: Chronology of Gully Development and Derivation of Meltwater from Latitude Dependent Mantle and Excavated Debris-Covered Glacier Deposits.

Abstract	4-1
Introduction	4-4
Geologic Setting	4-5
Crater Rays	4-6
Crater Age	4-7
Ice-Rich Mantling Deposits	4-8
Ice Content	4-9
Depositional History	4-9
Relative Stratigraphic Position	4-10
Ice-Rich Mantle & Gullies within the Host Crater	4-11
Theories of Gully Formation	4-12
Gasa Crater Interior	4-13
Gully Alcoves and Channels	4-14
Gully Fans	4-15
Evidence of Glacial Ice in the Host Crater	4-17
Glacial Accumulations in Craters	4-19
Comparison with Fresh Crater Zumba	4-22
Chronological Interpretation	4-23
Conclusions	4-26
Acknowledgments	4-27
References	4-28
Figures	4-45

Chapter 5: Decameter-Scale Pedestal Craters in the Tropics of Mars: Evidence for the Recent Presence of Very Young Regional Ice Deposits in Tharsis.

Abstract	5-1
Introduction	5-2
Observations	5-6
Decameter-scale Pedestal Craters	5-7
Chronological Constraints	5-8
Discussion	5-10
Interpretations and Conclusions	5-14
Acknowledgments	5-19
References	5-19
Figures	5-32

Chapter 6: Recent High-latitude Resurfacing by a Climate-related Latitude-dependent Mantle: Constraining Age of Emplacement from Counts of Small Craters.

Abstract	6-1
Introduction	6-2
Approach	6-6
Methods	6-8
Crater Count Data	6-9
Equatorial Rayed Craters	6-9
High-latitude Polygonalized Craters	6-11
Mid-latitude Craters	6-12
Interpretations of LDM Chronology	6-16
Conclusions	6-19
Acknowledgments	6-20

References	6-21
Figures	6-36

Chapter 7: Perspectives from Depositional Environments: Outstanding Questions and Exploration Strategy.

Introduction	7-1
The Stratigraphic & Sedimentary Perspective	7-3
Outstanding Questions in Martian Geoscience	7-4
Noachian Questions	7-5
Amazonian Questions	7-9
Is Mars Sample Return the Right Goal, Right Now?	7-13
References	7-18

CHAPTER 1

AN OVERFILLED LACUSTRINE SYSTEM AND PROGRADATIONAL
DELTA IN JEZERO CRATER, MARS: IMPLICATIONS FOR NOACHIAN
CLIMATE

(In Preparation as Schon, S.C., J.W. Head, and C.I. Fassett (2011) An
overfilled lacustrine system and progradational delta in Jezero Crater, Mars:
Implications for Noachian climate, *Planetary and Space Science*.)

Abstract

The presence of valley networks and open-basin lakes (those having both an inlet and outlet channel) in the late Noachian is cited as evidence for overland flow of liquid water and thus a climate on early Mars that might have supported precipitation and runoff. Outstanding questions center on the nature of such a climate, its duration and variability, and its cause. Open basin lakes, their interior morphology, and their associated channels provide evidence to address these questions. We synthesize the extensive knowledge of terrestrial open basin lakes, deltaic environments, and fluvial systems to assess these questions for Jezero crater, a 45 km diameter open basin lake and its 15,000 km² catchment area, ~645 km long drainage network, interior sedimentary facies, and ~50 km long outlet channel system.

Excellent exposures of sedimentary deposits at the mouth of the inlet channels permit us to address the question: Were these deposits formed in an alluvial fan or deltaic environment? We document the presence of extensive scroll bars and epsilon cross-bedding, both indicative of meandering distributary channels that are not observed on alluvial fans but are typical of fluvial-deltaic depositional environments. A fluvial-deltaic environment is further supported by the post-formational erosion of the deltaic complex: the present-day “delta front” is actually an erosional escarpment truncating delta plain features with the clay-rich prodelta environment, predicted from facies models to make up the outer third of the complex, having been largely removed by eolian erosion.

What was the nature of the lacustrine environment – was baselevel stable or did it fluctuate and perhaps cause the lake to undergo periodic desiccation? The extensive development via lateral accretion of scroll bars and epsilon cross-bedding, and the reconstructed sedimentary architecture suggests a stable baselevel, in contrast to an environment of constantly rising and falling baselevel related to variable input and evaporation that would favor vertical incision during lowstands. The development of the outlet channel is interpreted to have provided baselevel control in the Jezero open-basin lake. The maturity of the outlet channel (e.g., low-slope and presence of bar deposits), in contrast to the catastrophically scoured landscapes typical of dam-breach channels, favors a consistent overfilled hydrology for the

paleolacustrine environment. Was there a waning stage of activity during which lake level fell, the sedimentary deposits were incised and the locus of deposition migrated basinward? The observed deposits provide no indication of a waning stage or period. No incised channels are observed cutting the deposits. No evidence of a falling baselevel leading to forced progradation is observed. Therefore, we conclude that the hydrologically active period ended relatively abruptly, rather than over an extended period of time characterized by falling baselevel and incision.

What was the duration of overland flow, fluvial activity, and deposition -- could the deposits have formed on a decadal timescale, or was a longer period of time ($>10^5$ years) required? Sediment transport modeling studies of other valley network and related deposits on Mars have suggested durations in the decades to centuries range. We use a review of meander rates in terrestrial fluvial environments to provide a comparison for considering the temporal stability implied by the evolution of scroll bars; values of 20-40 years are not uncommon for the structure and migration implied by observations in Jezero. Taking sediment accumulation rates from a variety of terrestrial fluvial-lacustrine environments in conjunction with our estimates of the sedimentary basin-fill thickness suggest timescales of the order of $10^6 - 10^7$ years, far longer than implied by some sediment transport models, but still a short period of time geologically. The presence of significant residual accommodation space (space available for potential sediment accumulation)

in Jezero indicates that sediment transport into the lake terminated before the basin was completely filled.

What are the implications of the characteristics of the Jezero crater open-basin lake and fluvial system for Noachian climate? Climate conditions sufficient for sustained overland flow of water in the valley networks are required to fill Jezero crater, to cause its breaching in a non-catastrophic manner, and to form the significant fluvial-deltaic environment of laterally migrating fluvial channels and scroll bars formed at an apparently constant baselevel. The apparently stable baselevel is interpreted to mean that conditions during this period did not oscillate significantly. The lack of late-stage channel downcutting suggests that the conditions producing overland flow of water into the basin ended rather abruptly. Our estimates of the duration of fluvial activity (of order $10^6 - 10^7$ years) suggest longer times than previously suggested (years to centuries) by sediment transport models, but generally relatively short durations from a geologic perspective. Additional studies are necessary to assess whether these constraints apply to the rest of the valley network/open-basin lake population.

Introduction

Valley networks were first observed in Mariner 9 (Masursky, 1973) and Viking (e.g., Pieri, 1980; Carr and Clow, 1981; Carr, 1996; Carr, 2007) data of the 1970's. Interpreted as evidence of ancient fluvial erosion, the

degree to which these features are the result of precipitation (Craddock and Howard, 2002) or groundwater sapping (e.g., Laity and Malin, 1985) continues to be debated, with different minimum climatic requirements associated with each scenario. The resolution of this paleoclimate issue (e.g., Squyres and Kasting, 1994) – just how warm and wet, and for how long – has significant implications for Noachian hydrologic activity and potential habitability. Recent data suggest that periods of overland flow of liquid water occurred, evidenced by the aforementioned valley networks (e.g., Hynek and Phillips, 2001; Hynek and Phillips, 2003; Fassett and Head, 2008b; Hynek et al., 2010), lakes (e.g., Cabrol and Grin, 1999; Irwin et al., 2005; Fassett and Head, 2008a), and alteration mineralogies (e.g., Poulet et al., 2005). However, the cause, nature, and duration of the periods(s) has remained uncertain.

Well-preserved fluvial and lacustrine sedimentary deposits on Mars have been recognized in a variety of locations on the martian surface in the last fifteen years (e.g., Ori et al., 2000; Malin and Edgett, 2000; Malin and Edgett, 2003; Moore et al., 2003; Irwin et al., 2005; Fassett and Head, 2005; Weitz et al., 2006; Grant et al., 2008; Burr et al., 2009; Di Achille and Hynek, 2010). The depositional style and the nature of these deposits appear to range from alluvial fans to aggradational deltas, stepped-deltas, or progradational (Gilbert) deltas, and the depositional settings of particular deposits remain debated. The primary variable that differentiates these depositional styles is the stability and longevity of the alluvial/fluvial-lacustrine system required

for formation of the deposit. Much attention has focused on the Eberswalde crater deposit (formerly known as northeast Holden crater) (Malin and Edgett, 2003; Moore et al., 2003). Jerolmack et al. (2004) suggested that the deposit may not have a deltaic origin, but rather could have been formed by a riverine system without a standing body of water on a timescale of decades to centuries based on their modeling of alluvial-fan style development. In contrast, Bhattacharya et al. (2005) interpreted the deposit as resulting from a long-lived deltaic system based on evidence of multiple major channel avulsions and interpretation of a thick lacustrine section. Lewis and Aharonson (2006) proposed that rapid aggradational deposition of topset beds is suggested by shallowly dipping layers that they interpret as inconsistent with foreset bedding. This scenario implies multiple episodes of rising baselevel and is not consistent with the progradational interpretation of Wood (2006), which was based on evidence of several progradational lobes, their cross-cutting relationships, and multiple sinuous distributary channels in comparison to terrestrial analogs. In contrast to Eberswalde, Jezero crater has a defined outlet channel that creates the opportunity for a detailed analysis of the sedimentary construction of a martian “fan deposit” in an open basin environment (Fassett and Head, 2005).

In summary, the presence of valley networks and open-basin lakes in the late Noachian is cited as evidence for overland flow of liquid water and thus a climate on early Mars that might have supported precipitation and

runoff. Outstanding questions center on the nature of such a climate, its duration and variability, and its cause. Open basin lakes, their interior morphology, and their associated channels provide evidence to address these questions. Specifically, we summarize the terrestrial literature and address the following questions for the Jezero system: Were these deposits formed in an alluvial fan or deltaic environment? What was the nature of the lacustrine environment – was baselevel stable or did it fluctuate and perhaps cause the lake to undergo periodic desiccation? Was there a waning stage of activity during which lake level fell, the sedimentary deposits were incised, and the locus of deposition migrated basinward? What was the duration of overland flow, fluvial activity and deposition -- could the deposits have formed on a decadal timescale, or was a longer period of time ($>10^5$ years) required? What are the implications of the characteristics of the Jezero crater open-basin lake and fluvial system for Noachian climate?

In this contribution we start by reviewing a facies-based classification scheme for terrestrial lakes and identifications of paleolakes on Mars. Then we consider the Jezero system including the watershed, topography, accommodation space, and outlet channel. In section five, we (1) explicitly address distinctions between alluvial fans and deltas, (2) evaluate the depositional history of the Jezero deposits, and (3) outline evidence of their extensive post-depositional erosion. Then we turn to analysis of delta plain sedimentary structures including meanders and point bar sequences as well

as tabular channel sand bodies. In our discussion, we propose a scenario for the development and evolution of the Jezero paleolacustrine system from initial breaching of the crater rim to termination of hydrologic activity, which occurred prior to exhaustion of available accommodation space. Using terrestrial sedimentation rates and meander migration rates, we discuss plausible temporal constraints on this scenario. We also compare our estimates of formation time with recent sediment transport modeling studies of other sedimentary deposits on Mars that report extremely brief formation times (0.01 – 10 years; Kleinhans et al., 2010) and identify input parameters for these models that may require further refinement for application to Jezero and similar deposits. Finally, we conclude with potential constraints on the climate regime at this time and implications for the selection of future landing sites for Mars Science Laboratory (MSL) types of missions.

Classification of Lakes

Large terrestrial lakes are either of tectonic (e.g., East Africa rift lakes such as Tanganyika) or glacial (e.g., the Great Lakes of North America) origin (Johnson, 1984 and references therein). Tectonic lake assemblages, due to the relative paucity of ice ages, dominate the geologic record of lacustrine deposits. Terrestrial lake size parameters (depth, area, volume) do not correlate well with climate (precipitation/evaporation); in fact, a great diversity of lakes is found within particular climatic zones (Bohacs et al.,

2003 and references therein). Latitude, altitude, and drainage basin area also are not closely linked to lake size parameters. Rather, Bohacs et al. (2003) showed that lake volume, area, and depth have power-law distributions, which Fassett and Head (2008a) also documented for Mars paleolakes. These size distribution trends point to lakes as scale-invariant phenomena at least for moderate sized lakes (Meybeck, 1995; Turcotte, 1997; Bohacs et al., 2003; Fassett and Head, 2008a; Seekell and Pace, 2011).

Modern lakes are intricate biogeochemical systems with complex feedback mechanisms that have confounded the development of simple process-oriented lithofacies models. In contrast, the lacustrine rock record is separable into three distinct endmember lithofacies associations that together characterize most lacustrine basin fills (Figure 1). Such a tripartite division of lacustrine facies was first described by Bradley (1925) in characterizing the Green River Formation of the Uinta and Green River basins as composed of facies deposited in permanent freshwater lakes, in flooding and desiccating lakes, and lastly, under playa-like conditions. In extensive studies of Mesozoic-rift lacustrine strata of Newark Supergroup basins, Olsen (1990) identified and described a similar three-part division of lacustrine facies associations: “Richmond-type,” “Newark-type,” and “Fundy-type” (Olsen, 1990: his Figure 5). Richmond-type deposits are characterized by evidence of large depositional sequences; high relief sedimentary structures, such as prograding deltas, submarine channels, and turbidite

fans; and no indications of interspersed subaerial exposure or desiccation. Newark-type deposits show evidence of numerous significant changes in lake level that prevented the development of large sequence boundaries or high-relief sedimentary structures; however, these deposits contain repeated Van Houten sequences that are climatically keyed to Milankovich orbital cycles (Olsen, 1986). Finally, Fundy-type deposits are characterized by thin beds that record shallow perennial lake and playa-like conditions, and exhibit desiccation features, evaporites, and eolian dunes. A similar tripartite division of lacustrine facies associations (fluvial-lacustrine, fluctuating profundal, and evaporative) has been introduced by Carroll and Bohacs (1999) for more general application (Figure 1).

The Carroll and Bohacs (1999) nomenclature recognizes these associations as endmember lithofacies associations characterized from a very large suite of ancient and modern lacustrine systems (Bohacs et al., 2000; Bohacs et al., 2003). These endmembers are distinctive in their lithology, sedimentary structures, and biogeochemistry, but are relatively independent of age, water depth, and thickness (Bohacs et al., 2000; Bohacs et al., 2003). What, then, controls the occurrence of these lacustrine facies? Consideration of many lacustrine process-response relationships affecting sediment delivery and dispersal led to the development of a predictive classification scheme of lake basins as overfilled, balance-filled, or underfilled (Carroll and Bohacs, 1995, 1999). These lake-types (Figure 1) are controlled by two primary

factors: accommodation space (related to geologic and tectonic setting) and the supply of water and sediment (related to climate). While modern lakes are complex systems, these simple controls have significant explanatory power for lacustrine lithofacies records at thicknesses greater than ~ 1 m, and therefore provide a predictive framework for the development of lacustrine basin fills (Figure 1). The strong association of the endmember lithofacies with the lake classification scheme allows for prediction of lake type based upon limited outcrop data and sedimentary structures. Lake type and facies distribution predictions of this kind have proven to be a very effective framework for interpreting lacustrine basin fills (e.g., Johnson and Graham, 2004; Bohacs, 2004; Keighley, 2008). In the present study, we apply this framework to Mars and show that Jezero was an overfilled lake system. However, first we turn to the record of lakes on Mars and their identification to show that Jezero is not unique in its structure or watershed.

Lakes on Mars

Noachian-age paleolakes were first identified with Viking imagery (Goldspiel and Squyres, 1991). Additional potential paleolakes were identified by De Hon (1992), Forsythe and Blackwelder (1998), Cabrol and Grin (1999, 2001), and Mangold and Ansan (2006), as well as Irwin et al. (2002, 2004) who described the very large Eridania basin associated with Ma'adim Vallis. Because lake basins are identified based upon topographic

relations (Hutchinson, 1957; Wetzel, 2001), the Mars Orbiter Laser Altimeter (Smith et al., 1999) and digital terrain models derived from the High Resolution Stereo Camera (Neukum et al., 2004) have provided additional crucial data for discerning potential paleolakes. Using these topographic datasets in conjunction with multi-mission visual images, Fassett and Head (2008a) catalogued 210 open-basins with distinct inflowing valley networks and outlets. The vast majority of these basins are impact crater related. Many intra-valley paleolakes are preserved because of the relatively immature (e.g., Gutiérrez, 2005) martian landscape (e.g., Stepinski et al., 2004).

How have subsequent geologic processes affected the paleolakes since their formation? Mars paleolakes are inferred to be Noachian in age (>3.6 Ga) based upon relations to the valley networks that sourced them (Carr, 1996, 2007; Fassett and Head, 2008b), but this does not reflect the full geologic history of these basins. Significant erosion and modification have occurred, such as continued infilling and resurfacing of the basin interiors. In their survey work, Fassett and Head (2008a) noted that 50% of the basins cataloged contained clear evidence of volcanic resurfacing. In addition to volcanic resurfacing -- primarily by Hesperian ridged plains (e.g., Scott and Tanaka, 1986; Greeley and Guest, 1987; Head et al., 2006) -- impact crater ejecta (e.g., Cohen, 2006), volcanic tephra (e.g., Wilson and Head, 2007), eolian sediments (e.g., Fenton et al., 2003), and glacial deposits (Goudge et

al., 2011) can be important post-lacustrine basin fills. Therefore, while paleolakes are relatively common, and distributed throughout the southern highlands, preserved and observable sedimentary features associated with these basins are relatively rare due to post-lacustrine erosion, subsequent non-lacustrine basin fills, and resurfacing (Goudge et al., 2011). Well preserved basins with clear sedimentary deposits associated with valley networks are likely to number no more than 40 to 60 globally based upon recent surveys (Irwin et al., 2005; Fassett and Head, 2008a; Di Achille et al. 2008; Di Achille and Hynek, 2010). Consequently, while the basin and watershed themselves are not unique in their area and volume relationships (Fassett and Head, 2008a) the excellent sedimentary exposures in Jezero crater make this system particularly attractive for detailed investigation.

Jezero Lacustrine System

Jezero crater (18.4°N, 77.7°E) is a 45-km diameter impact crater located in the Nili Fossae region of Mars. Fassett and Head (2005) mapped the associated valley networks, which drain a 15,000-km² watershed (Figure 2), and identified two sedimentary fans in the basin that we interpret as a single sedimentary assemblage (Figure 3). The watershed and surrounding Nili Fossae region are a mineralogically diverse Noachian terrane where many aqueous alteration products -- such as phyllosilicate clays and carbonates -- have been detected by visible/near-infrared spectroscopy

(Bibring et al., 2006; Mangold et al., 2007; Ehlmann et al., 2008a; Mustard et al., 2008). In addition to the aqueous alteration minerals detected in the watershed, phyllosilicate and carbonate detections within Jezero crater sedimentary deposits suggest that these sediments were transported from the watershed rather than weathered in place (Ehlmann et al., 2008b; Murchie et al., 2009).

The watershed ranges in elevation (relative to the Mars' datum) from ~250 m along the northern drainage divide to -2400 m, the elevation of the valley entrances and the outlet. The valley networks are composed of 645 km of mapped channels (Figure 2). While the Strahler order (number of tributaries upstream) is low (third-order), the main valleys are quite mature. They are low slope (0.5 and 0.7 degrees in their lower reaches) and have meanders that are incised hundreds of meters (Figure 2, Figure 3). Low drainage densities (by terrestrial standards) are the norm for even the most well developed Noachian valley networks (Baker and Partridge, 1986; Carr and Chuang, 1997; Hynek and Phillips, 2003); the Jezero watershed is not unusual (0.043 km^{-1}). The lack of observable high-order tributaries on Mars leading to commensurately less mature landscapes has been explained by various mechanisms, including impact gardening which could remove rills and small tributaries (e.g., Hartmann et al., 2001), high infiltration rates which could minimize overland flow in tributaries (e.g., Carr and Malin,

2000), and a shorter period of hydrologic activity during which erosion in tributaries was modest (e.g., Stepinski et al., 2004).

Accommodation space (the space available for potential sediment accumulation) in the Jezero basin was created by the impact event during the Noachian that excavated a 45-km diameter crater. Complex craters, such as Jezero, are characterized by broad, level floors, and often have terraces (circumferential rim failures), and central peak elements. Systematic crater depth-diameter ratio trends have been investigated by Garvin et al. (2003), and we use these to estimate original topographic profiles and the maximum accommodation space for water and sediments within Jezero (Figure 4). A fresh crater of similar size is also used for comparison.

Using MOLA topography to characterize more than 6,000 impact craters, Garvin et al. (2003) systematically investigated the depth-diameter relationship for Mars impact craters and developed a refined power-law relationship for complex craters: $d = 0.36D^{0.49}$ where d is crater depth (km) and D is diameter (km). The Garvin et al. (2003) data predict a depth (d/D) of 2320 m (0.0516). Observed from MOLA profiles, the actual depth (d/D) of Jezero is 1080 m (0.0241). A comparable fresh crater (125.75°E, 8.25°N) selected from the crater catalog of Barlow (1988) has an actual depth of 1960 m (0.0435). The profoundly shallower profile for Jezero compared to statistical relations for complex craters (e.g., Garvin et al., 2003) and a similarly sized fresh crater (Figure 4), show that Jezero has experienced

substantial filling, ~1 km (Ehlmann et al., 2008b). Models of ejecta thickness decay by Cohen (2006) suggest that at most 24 m of this fill is ejecta from subsequent impact craters.

The present basin interior is covered by a thin volcanic unit that is observed to embay the fan deposits (Figure 5). Near the fan deposits we estimate this material is no more than 10-30 m in thickness based upon topographic relationships observed at eroded embayment contacts (Figure 5). The crater size-frequency distribution ($n=724$) observed on 344-km² of this unit suggests an Early Amazonian age of 1.4 Ga using Hartmann (2005) isochrons (Figure 6). While the volcanic unit is pervasive as a cap unit on the central basin floor, sedimentary “windows” are observed in relation to the fan deposits 10.5 km from the crater rim (e.g., Figure 5). These windows occur where previously high-standing sedimentary material was embayed by the volcanic unit. Subsequently, the sedimentary material has further eroded, leaving the more resistant volcanic material as a raised rim around a depression of the sedimentary material (Figure 5) with abundant dunes of reworked deltaic material. At these sedimentary windows, the known depth-diameter relationships described above enable us to estimate a thickness of ~750 m for the basin fill.

Baselevel within the paleolake was controlled by the outlet channel on the east side of the crater. Development of the outlet channel originated from initial overtopping of the crater rim and subsequent erosion of the rim breach

(250-300 m) to the present condition where the breach and fan deposits share a topographic contour within a few tens of meters (see line on Figure 4). The watershed/basin area ratio of $\sim 10:1$ implies, for example, runoff production from the watershed of ~ 10 cm/yr (equivalent to an arid terrestrial environment, Köppen classification BW) and a discharge of $50 \text{ m}^3/\text{s}$ to balance 1 m/yr of evaporation from the lake. While Noachian evaporation rates are uncertain (Irwin et al., 2007), this scenario illustrates a plausible minimum level of activity (e.g., Andrews-Hanna and Lewis, 2011) that development of the outlet channel indicates was exceeded.

Inflow to Jezero in excess of evaporation could have occurred and been lost by infiltration to a regional groundwater system, or discharged from the basin via the outlet channel, or both. Loss to regional aquifers, if any, is speculative and difficult to constrain from remote data. The outlet channel is mapped for ~ 53 km before it is obscured by overlying geologic units (Figure 2). Incision of the breach and the maturity of the channel (Figure 7) indicate that an overflowing hydrology developed. Along its course, the outlet channel meanders and fluvial sedimentary deposits are preserved (Figure 7). Planar bedding within light-toned material is common on the inside of meander bends and along reaches of the channel (Figure 7). The channel itself must have eroded the sediment for these features because the 40 km distance across the basin from the inflowing valley networks would have effectively trapped all sediment from the watershed. The development of this mature,

low slope (0.6°) outlet channel requires that Jezero was a stable overfilled paleolake (Figure 1). The stable baselevel near -2400 m controlled by the outlet (Figure 4) would be an ideal lacustrine environment for the development of a progradational delta (Figure 1).

Deltaic Deposits

The continued ambiguity regarding the depositional history of fan deposits on Mars has led to a complex collection of terms used in reference to these sedimentary features, including “fans,” “alluvial fans,” “delta fans,” “alluvial deltas,” “distributary fans,” and “deltas.” This issue of uncertain depositional conditions was raised directly by Malin and Edgett (2003) and has not been adequately resolved (see discussion in Moore and Howard, 2005). However, new sub-meter resolution data from the HiRISE camera on the Mars Reconnaissance Orbiter (McEwen et al. 2007) enable a detailed assessment of these sedimentary deposits. In conjunction with a comparison to terrestrial sediment transport and deposition processes, this allows for firm discrimination between an alluvial fan origin and deltaic origin for the sedimentary assemblage in Jezero. We present evidence of meandering distributaries on the Jezero crater fan deposit that indicate these deposits are of fluvial-deltaic origin and contrast their distinguishing features with the defining attributes of alluvial fans as well as sediment deposited under unstable lacustrine conditions.

There are several very fundamental differences in the formation of alluvial fan deposits compared to deltaic deposits. Alluvial fans are entirely subaerial, semicircular deposits that radiate from sharp breaks in slope, most commonly along mountain fronts. They are deposited primarily in stream floods, sheet floods, and debris flows, resulting from vigorous but episodic precipitation events. Alluvial fan streams are braided and modestly entrenched; channel cuts and fills are common in alluvial fan stratigraphy and sorting is typically poor because transport distances are short (Blissenbach, 1954; Bull, 1977; Harvey et al., 2005).

In contrast, deltas are partially subaerial sediment masses that are deposited where a low gradient channel debouches into a standing body of water. Changes in baselevel (i.e., local sea level or lake level) exert an important control on delta morphology by shifting depositional trends (see discussions in Payton, 1977; Catuneanu, 2006; and references therein). However, primary delta morphology is controlled by the rate of sediment input relative to reworking or removal by energy sources within the basin (Galloway, 1975). This characteristic has led to a ternary classification scheme for delta morphologies, which contrasts the relative influences of sediment supply, wave energy, and tidal variations (Figure 8).

The demarcation of alluvial fans and deltas, based upon extensive terrestrial field studies (e.g., Bull, 1977), illustrates distinct contrasts in gross morphology that are applicable to martian and remote sensing studies

of deltas (e.g., Pondrelli et al., 2005; Pondrelli et al., 2008; Hauber et al., 2009; Farris, 2009) and alluvial fans (e.g., Moore and Howard, 2005; Kraal et al., 2008a; Williams and Malin, 2008; Hardgrove et al., 2009; Hardgrove et al., 2010). Alluvial fans are semi-conical in shape, restricted in radial length, convex in cross-profile, and have high values of radial slope. In contrast, deltas have generally lobate planforms and have low radial and cross-profile slopes (Blair and McPherson, 1994). Deltas are well-sorted fine-grained deposits, compared to the coarser, poorly sorted, sediments that construct alluvial fans. Alluvial fan sediments are sourced from smaller drainage basins on bold topography and transported shorter distances by high-competence streams or debris flows (Blair, 1999). Contrarily, deltaic sediments are typically suspended load and bedload of extended river systems. The higher slope of alluvial fans ($>1.5^\circ$) leads to flows that are often supercritical (e.g., sheetfloods), while flows in low-slope delta environments are subcritical. The conical form of alluvial fans leads to rapid expansion and attenuation of flow, which therefore reduces the competence and capacity of the stream, leading to rapid sedimentation near the fan apex (Blair and McPherson, 1994). In contrast, deltas commonly have leveed channels, meandering distributaries, overbank deposits, splays, and abundant associated channel and mouth bars (e.g., Coleman, 1981).

Deltas are divided into three environments of deposition (Rich, 1951): the delta plain, the delta front, and the prodelta. The delta plain is comprised

of alluvial sediments and includes meandering distributaries' floodplains, marshes, and beach environments. The more inclined ($\sim 5\text{-}7^\circ$) delta front is the primary locus of deposition, while the distal prodelta receives the finest sediment fraction. These environments are associated with topset (delta plain), foreset (delta front), and bottomset (prodelta) beds of large-scale prograding clinoforms in seismic reflection data. Generally, depositional angles are quite low even in foreset beds (Mitchum et al., 1977), which effectively prevent the recognition of such large depositional packages in martian remote sensing studies. At present, the front of the Jezero fan is a steep ($\geq 10\text{-}30^\circ$) erosional escarpment, not a primary depositional feature.

The resurfacing history of the basin provides significant evidence that the Jezero delta was substantially larger and has experienced significant erosion prior to the most recent resurfacing of the present basin floor. The most recent resurfacing event has been dated to the Early Amazonian (best fit: 1.4 Ga) based upon the size-frequency distribution of superposed craters (Figure 6). The erodibility of the deltaic deposits and the relative strength of the embaying volcanic unit are shown in Figure 5, where the resurfacing unit can be seen surrounding what is now an eroded depression. This sedimentary window was a positive topographic feature when the embaying unit was emplaced, but has since eroded somewhat further. Dunes of the eroded sediment are present along the contact. Differential crater preservation on the contact between these units occurs where a portion of a crater is well

preserved on the embaying unit, but the remainder of the crater that would have been impinging the sedimentary material has been removed by erosion, further demonstrating the post-depositional erosional retreat of the deltaic deposits. Isolated distal remnants of sedimentary material, located ~ 3 km from the continuous deposit, rise ~150 meters above the basin floor and also serve as indicators of the larger previous extent of the delta (Figure 9). These isolated kīpuka-like remnants are entirely embayed by the resurfacing unit and, with peaks approximately 50 m below the height of the present escarpment, represent a minimum previous extent of the delta nearly twice as large as the continuous fan deposit of today (Figure 9).

Post-lacustrine erosion of the delta and resurfacing of the basin floor obscure the prodelta region from current observation (prodelta would be basinward and at lower elevation than the most distal sedimentary deposits that are observable). The present “delta front” is an erosional feature – an escarpment – resulting from post-depositional erosion of the deposits and is not related to primary deposition. Rather, sedimentary structures characteristic of a delta plain environment are truncated by the scarp (Figure 10). Therefore, the delta plain environment, particularly the western portion, remains as the most well-preserved, and depositionally-representative, portion of the Jezero delta complex (Figure 10) and is the focus of the next section.

Delta Plain Sedimentary Features

Meanders and point bar sequences are well-studied features of both the Quaternary and older geologic record because of their importance to petroleum systems (e.g., Smith 1988; Bridge and Tye, 2000), navigation (e.g., Fisk, 1947), and natural hazard management (e.g., Johnson, 2005). In alluvial stream systems, the active channel morphology is controlled by the interaction of flow on boundary materials that have been deposited by the stream and can be eroded and transported by the stream. In this environment, meanders form naturally as a result of secondary spiral currents that enhance flow velocity and channel depth along the outer margin of a bend (Leopold and Wolman, 1960; Ikeda et al., 1981; Blondeaux and Seminara, 1985; Parker and Andrews, 1986; Ikeda and Parker, 1989; Stølum, 1996; Seminara, 2006; Howard, 2009). The evolution of meanders was studied on numerous alluvial streams by Brice (1974), who devised a canonical classification scheme for meander loops based on their degree of symmetry and geometric complexity. The natural tendency of meanders is to increase the sinuosity of the channel system by eroding their outer banks and depositing sediment along their inner banks where point bars develop. They also translate downstream. Variations in streamflow, sediment load, and relative proportions of washload and bedload influence meander development and behavior (e.g., Schumm, 1963).

Point bars (e.g., Nanson, 1980) are prograding, diachronous, time-transgressive, laterally continuous, fining upward sequences that form at the inner bank of meanders (Figure 11). The growth of point bars produces distinctive lateral accretion topography (Figure 11) characterized by scroll bars and intervening swales (e.g., Puigdefabregas, 1973; Hickin, 1974; Hickin and Nanson, 1975; Nami, 1976; Schumm, 1985). The planimetric signature of lateral accretion topography is easily recognized in remote sensing data due to the distinctiveness of the scroll bar pattern (Figure 12). First studied in detail by Fisk (1947) on the lower reaches of the Mississippi River, point bars form in meandering systems of all scales (Smith, 1998) and are well studied sedimentologically (e.g., Allen, 1965). Recently, so-called counterpoint bars have also been described (Smith et al., 2009). These deposits develop downstream from point bars at the point of meander inflection and thicken distally from the upstream point bar. While point bars are sand-dominated, counter point bars (also called concave bank-bench deposits) are predominately composed of silt (Smith et al., 2009). On the Jezero delta plain, it is possible that erosion products from such counterpoint bar deposits could contribute to the detections of clay minerals (Ehlmann et al., 2008b), but counterpoint bar deposits are not observed directly.

In addition to the distinctive topographic signature of point bars (scroll bars, Figure 12), prograding point bars also develop inclined accretion surfaces, termed epsilon cross-bedding (Allen, 1963), that are visible in cross-

section (e.g., Nami and Leeder, 1978; Stewart, 1983; Smith, 1987). The paleocurrent direction is parallel to the strike of the inclined accretion surfaces. The inclined lateral accretion surfaces dip in the direction of channel migration (arrows in Figure 13). A one-km diameter crater superposed on the western portion of the delta plain provides the necessary exposure and reveals epsilon cross-bedding in its walls (Figure 13). The inclined lateral accretion surfaces indicate that sediment was extensively reworked at this location as a succession of point bars prograded in multiple directions (Figure 13).

Discussion

The facies distinctions and lake classification scheme developed by Bradley (1925), Olsen (1990), Carroll and Bohacs (1999), and Bohacs et al. (2000) provide a powerful interpretative framework (Figure 1) for understanding the development of lacustrine basin fills. On Mars, the accommodation space dimension of this framework is simplified because most paleolakes, including Jezero, occur in impact craters. Because impact cratering is a well-understood process (e.g., Melosh, 1989; Barlow, 2009, and references therein), estimates can be made of the basin fill (Figure 4). The somewhat dendritic valley networks that drain the watershed and source the fan deposits are consistent with precipitation (e.g., Fassett and Head, 2008a; Fassett and Head, 2008b). The existence of an outlet channel from Jezero

crater (Figure 7), similar to many open-basin paleolakes, indicates that this basin had an overflowing hydrology for some length of time. The outlet channel has a meandering planform with associated bar deposits (Figure 7) that indicate formation under stable discharge. In contrast, singular dam-breach-flood events scour the substrate and do not develop similar bars (e.g., Baker and Milton, 1974; Rydlund, 2006; Lamb et al., 2008).

While most Noachian paleolakes have been extensively resurfaced (e.g., Fassett and Head, 2008a; Goudge et al., 2011), the Jezero system contains well-exposed sedimentary deposits. The coincident elevation of the outlet channel notch and the surface of the fan deposits (Figure 2) indicate that the sedimentary assemblage (Figure 3) developed via deltaic progradation rather than as a deepwater submarine fan (e.g., Bouma et al., 1985), or as an alluvial fan in a playa like environment (e.g., Blair, 1999; Hardgrove et al., 2010). Extensive erosion of the fan deposits occurred prior to the most recent resurfacing of the basin interior (Figures 9 and 10), dated to the Early Amazonian (~1.4 Ga), and subsequent erosion has continued to alter the deposits (Figure 5). While the sharp erosional relief of the delta front scarp is the most obvious manifestation of the erosional history, impacts and eolian erosion have also altered the local topography of the delta plain and improved exposure of some depositional sedimentary structures. Therefore, the delta plain environment provides the most extensive geological

evidence for constraining depositional processes and interpreting the lacustrine system.

An overfilled lacustrine system: The Jezero fan deposits (i.e. delta plain) are very low slope ($\sim 0.5^\circ$) and approximately of the same contour (~ 2400 m) as the outlet channel that controlled baselevel. Detections of phyllosilicates (Ehlmann et al. 2008b) occur in areas of swale-and-ridge topography that we interpret as scroll bars (Figure 12). Scroll bars are evidence of lateral accretion and point bar sequences deposited by meandering channels (Figure 11). The sediment cohesion required to form these features (e.g., Peakall et al., 2007) is interpreted to be provided by the clays that have been detected from orbit (Ehlmann et al., 2008b). Spectral identification of clays may also be attributable to the co-development of silt-rich counterpoint bars (e.g., Smith et al., 2009) in association with the sand-dominated point bars responsible for the scroll bars.

Truncations within these scroll bar patterns suggest extensive reworking of alluvial sediment by meandering channels (Figure 12). A large (~ 1 km-diameter) crater that postdates the depositional epoch provides cross-sectional views of the deposits. Multiple generations of point bar sequences deposited by meandering channels are required to account for the multiple sets of epsilon cross bedding (Allen, 1963) observed within the crater walls (Figure 13). Elongate sediment bodies observed on the delta plain surface are interpreted as channel sands (green lines on Figure 14). These

stratigraphically higher (Figure 15) channel systems fed more distal depocenters of the delta complex that have subsequently been eroded. Finer grained overbank and splay deposits from these channels are likely to represent a majority of the material eroded from the fan surface between these distinct channel sand bodies (Figure 14, Figure 15B). Cross-cutting relationships between the channel sand bodies (Figure 14) suggest that some of these deposits were emplaced in a series of sequential depositional episodes (e.g., lobe/channel-switching). The channel sand bodies are more resistant to erosion and as expected compositionally, phyllosilicate detections are not associated with these features (Ehlmann et al., 2008b). Channel sand bodies are common in terrestrial deltaic systems in a variety of physiographic settings (Busch, 1971; LeBlanc, 1972; Busch and Link, 1985). Similar high standing channel sand bodies have been documented in fluvial sandstone formations of the Colorado Plateau (e.g., Stokes, 1961). The channel sands (Figure 14) are consistent with our interpretation of a previously more extensive delta (e.g., Figure 10).

Utilizing the facies and basin classification framework (Figure 1), we interpret Jezero as an overfilled basin. Initial accommodation space is the result of the formation of a Noachian-aged fresh impact crater. Crater degradation processes and precipitation-fed valley networks breached the crater rim and initiated filling of the basin. Crater breaching by valley networks is common on Mars (Fassett and Head, 2008a; Enns et al., 2010)

and may be aided by infiltration through impact-related faults (e.g., Kumar and Kring, 2008; Kumar et al., 2010) akin to infiltration and piping induced dam failures (Bedmar and Araguas, 2002; Richards and Reddy, 2007), or by topographic ponding and subsequent overtopping and down-cutting of the crater rim. Once these valley networks flooded the crater basin, formation of the outlet channel began.

Sediments transported by the inflowing valley networks constitute the fluvial-lacustrine facies association that in our interpretation is a majority of the basin fill (e.g., Figure 4). Episodic filling and desiccation of the basin (e.g., fluctuating-profundal facies) is inconsistent with the large high-relief deposits (Figure 3) and the mature outlet channel (Figure 7). If Jezero was a dominantly balanced-fill lake, the outlet channel would be absent or very immature. Any outlet would be relatively short and dominated by scour rather than exhibit the bar deposits that are observed (Figure 7). In a dominantly balanced-fill lake, the locus of sediment deposition would have experienced major shifts shoreward (highstand systems tract) and basinward (lowstand systems tract) with the fluctuating lake level. Transgressive surfaces would be dominant features and lateral accretion deposits (Figures 12 and 13) would not have developed. Desiccating lowstands would lead to channel incision (e.g., Weitz et al., 2006, her Fig. 4c) and possibly even alluvial fan deposition.

Similarly, a high-discharge cataclysmic filling of the basin, breaching of the rim, and near-immediate decline of the inflowing channels is not consistent with the observed deposits in our interpretation. Such a high discharge, short period, scenario would not result in the stable baselevel required to form the lateral accretion deposits that are observed. Depositional features within the outlet channel require a sustained discharge for sediment erosion, transport, and deposition (Figure 7). Because the basin would be an excellent sediment trap, sediment in the outlet channel must be eroded by the outlet channel – a singular catastrophic outflow would only scour a path (e.g., Rydlund, 2006); the outlet channel morphology requires a stable overfilled lacustrine hydrology.

Sedimentation rates: Measurements of sediment package thickness, in conjunction with terrestrial experience with sedimentation rates, allow us to estimate minimum durations of stable activity in the Jezero system. Comparing the topographic profile of the present Jezero crater with a fresh crater of the same diameter and depth-diameter relationships (Garvin et al., 2003) suggests a likely lacustrine basin fill of ~750 m. Even the most conservative estimate of sediment thickness, taken by differencing the elevation between the most basinward and most proximal present-day sedimentary exposures, would suggest a thickness of ~300 m. Terrestrially, impact crater basin analogs are rare due to the Earth's much more vigorous geologic history. Lake El'gygytgyn in the Russian arctic (67.5°N, 172°E) is an

open basin lake system formed in a 12-km diameter Pliocene-age crater. Studies of shallow sediment cores and seismic imaging of the sediment fill suggest deposition rates of 3-12 cm / Kyr (Melles et al., 2005).

However, sediment accumulation rates have been observed or measured extensively in a variety of other terrestrial basin environments. Therefore, conservative assumptions derived from a very large dataset enable estimates of the minimum timescale for deposition. In a seminal study, Sadler (1981) compiled nearly 25,000 measures of sediment accumulation rates and showed conclusively that such rates are inversely related to the time span of the measurement. For lacustrine environments, the Sadler (1981) rates span four orders of magnitude ($\sim 0.01 - 100$ m / Kyr) depending on the time span of the measurement ($10 - 10^8$ yr). However, sediment accumulation is not uniform in a basin. In particular, a progradational delta system has localized depocenters and sediment deposition is concentrated along the delta front (e.g., Corbett et al., 2006). Assuming that a delta front sediment accumulation rate of cm per annum (Hori and Saito, 2008) prevailed for the entire depositional history of Jezero suggests a lifetime on the order of 10^4 (Earth) years for the system, while median paleolacustrine sediment accumulation rates of Stadler (1981) indicate a potential lifetime of $10^6 - 10^7$ years. An important consideration in such calculations is the availability of sediment supply in the watershed. Terrestrially, biota is an important component of chemical and physical weathering processes, but

biota can also retard erosion (Dietrich and Perron, 2006); the net effect of the absence of these counteracting influences on martian sediment generation and transport is unknown. Impact gardening is likely to be a dominant sediment-forming process on Mars today (e.g., Hartmann et al., 2001), but rates of impact gardening during the Noachian are poorly known. The higher impact flux in the Noachian would suggest more rapid impact gardening than at present, but an early thick atmosphere could have shielded the surface from small impacts, reducing the efficiency of gardening (e.g., Hartmann and Engel, 1994).

Measurements of the clearly identifiable scroll bars (Figure 12), in conjunction with terrestrial experience with meander migration rates, can also suggest minimum durations of depositional activity in the Jezero system. Scroll bars here (Figure 12), elsewhere on Mars (e.g., Eberswalde; Wood, 2006), and terrestrially (e.g., Hickin and Nanson, 1975; Schumm, 1985;) commonly exhibit numerous cutoffs and unconformities formed by erosion and redeposition of previously laterally accreted sediments by an active channel. Cutoff events have an important dynamical influence on the continued evolution of other meanders (Camporeale et al., 2008; Constantine and Dunne, 2008). Therefore, lateral measurements of continuous point bar deposits are very conservative estimates for the overall lifetime of the system. In Jezero these individual features are commonly tens to hundreds of meters in width (Figure 12).

The development of meander loops (Rich, 1914) and their evolution and migration through time has long been an empirical inquiry of geologists. As products of the fluvial environment, the formation and evolution of meanders and their geological preservation in point bars are affected by the major factors controlling alluvial stream channel form. Seminal field studies by Brice (1974), Leopold et al. (1964), Schumm (1985), and others have suggested relations between stream parameters and meander bend migration.

We utilize a large dataset of stream bank meanders that was originally compiled for the Transportation Research Board (TRB) of the National Academies for purposes of civil engineering (Lagasse et al., 2004). This dataset encompasses 89 rivers in the continental United States, and includes data from 1,503 unique meander bends at 141 field sites. These study sites were re-occupied multiple times allowing for comparison of the meanders over years and decades. Some meanders were cutoff, while at other locations, new meanders were observed to form along previously straight reaches. We employed a variety of historical imagery for reconnaissance of each meander bend and excluded all field sites impacted by artificial revetments (e.g., riprap and other bank stabilizations) or channel modification (e.g., sand and gravel mining). Excluding meander cut-offs, 1,009 measurements of meander migration were calculated with a minimum (0.03 m/yr), mean (3.76 m/yr), median (2.52 m/yr), and maximum (30.00 m/yr) rate of meander migration

(Schon et al., in preparation). These rates suggest that plausible timescales for the formation of the individual scroll bar sets (Figure 12) are likely on the order of decades (~20-40 years) assuming average terrestrial migration rates.

Comparison to modeling results: These estimates of the duration of lacustrine activity in Jezero ($\sim 10^6 - 10^7$ years) are substantially longer than the short minimum timescales of formation (0.01 – 10 years) calculated for some Mars fan deposits using sediment transport models (Kleinhans, 2005; Kleinhans et al., 2010). Neither Kleinhans (2005) nor Kleinhans et al. (2010) considers the Jezero system specifically. But, for example, modeled minimum timescales for the sedimentary fan deposits of Ma'adim, Nanedi, Sabrina, and Hypanis valleys are all less than a decade. While the sediment transport mechanics employed by Kleinhans (2005) and Kleinhans et al. (2010) are well-validated, multiple geologic features of the Jezero system and different assumptions explain our divergent conclusions about the length of activity inferred: (1) while discharge is difficult to constrain with certainty, we suggest that bank-full discharge is not a good approximation for a long-lived fluvial system such as Jezero; (2) the valley networks may have been detachment-, or supply-limited, akin to a bedrock river; and (3) the Jezero delta plain morphology requires a cohesive component (clay; Ehlmann et al., 2008b) and sand, which are inconsistent with Kleinhans et al. (2010) assumptions regarding the size-distribution of transported clasts (“a median grain size of 0.1 m and a 90th percentile size of 0.6 m diameter”). Meandering

systems transporting primarily gravel and larger sediment are unknown; these materials give rise to braided channel patterns terrestrially (Harms et al., 1975).

Kraal et al. (2008b) suggest that so-called “stepped deltas” formed quickly (years to decades) due to enormous discharges and fast-rising lake levels. In contrast, the deltaic architecture at Jezero suggests a stable baselevel and a longer time-scale for formation. While Jezero has a large watershed drained by established valley networks, the Kraal et al. (2008b) example has an extremely short channel (~20 km) that is implied to have had a discharge comparable to the Rhine or Mississippi Rivers in their analysis. Kraal et al. (2008b) favor a large release of groundwater rather than a precipitation-fed origin for the discharge. Therefore, these “stepped delta” deposits (see also, Weitz et al., 2006) are indicative of local or regional groundwater releases (Kraal et al., 2008b) in contrast to progradational deltaic deposits such as Jezero or Eberswalde (e.g., Bhattacharya, 2005; Wood, 2006; Pondrelli et al., 2008; Dietrich, 2010) that due to their large catchments are more representative of general climatic conditions during their deposition (e.g., Di Achille and Hynek, 2010).

Conclusions

The Jezero crater open-basin paleolake contains eroded deltaic deposits (Figure 3). While the apparent modern delta front is actually an

erosional scarp (Figure 10), the delta plain is well exposed with an extensive pattern of scroll bars (Figure 12) that formed via lateral accretion from meandering distributary channels (Figure 13) coincident with phyllosilicate detections (Ehlmann et al., 2008b). The presence of a stable subaerial delta plain environment with meandering distributaries indicates that the delta prograded during an extended period of baselevel stability in an overfilled lake system (e.g., Figure 1). Isolated remnants of the delta (Figure 9), embayment and erosion relationships (Figure 5), and stratigraphically higher elongate channel sand bodies (Figure 14, Figure 15) point to the larger previous extent of the delta. The larger previous extent of the delta is evidence of extensive post-depositional erosion predominantly prior to Early Amazonian volcanic resurfacing of the basin floor (Figure 5). Sedimentary bars in the outlet channel (Figure 7) provide independent evidence of the stable overfilled lacustrine system. Topographic comparison with fresh craters and impact crater scaling relationships (Figure 4) require a significant basin-fill in Jezero consistent with the delta and lacustrine sedimentation.

Our interpretations of the depositional characteristics and plausible sedimentation rates suggest that formation of the Jezero delta required a period of $\sim 10^6 - 10^7$ years to form. This suggests a minimum persistence of climatic conditions sufficient for valley network formation and open basin hydrology during a portion of the Noachian. Our analysis has not revealed

any evidence of waning-stage channel incision or forced progradation, indicating that the end of the Noachian climate period during which these deposits formed was likely to have been rapid. The accommodation space that remains (200-300 m) indicates that while our interpretation implies that these environmental conditions persisted in the Noachian for longer than required by some models (e.g., Kleinhans, 2005; Kraal et al., 2008b), sediment transport and deposition in the Jezero paleolacustrine system was fundamentally limited by a secular shift in climate.

The Jezero delta is an attractive target for in situ exploration by missions following MSL (Golombek et al., 2011; Grant et al., 2011) that will seek to characterize the stability and longevity of Noachian habitable environments. Based on analysis of Jezero during the MSL landing site selection process, this would likely require improvements in precision landing capabilities (reduction in landing ellipse size) over the MSL architecture. Observations from the crater floor of the deltaic escarpment would reveal the finest-scale details of the bedding, while the traversable terrain of the delta plain contains sedimentary structures and mineralogical diversity deserving further investigation (e.g., Fassett et al., 2007; Ehlmann et al., 2008b). Finally, the sedimentary basin fill within Jezero presents a compelling target for future subsurface exploration with capabilities being developed by NASA's Mars Technology Program (Miller et al., 2004). Terrestrial overfilled lacustrine systems contain numerous organic-rich units with the potential for

excellent biomarker preservation. If drilled, the Jezero sedimentary record could elucidate the history of Mars' climate and surface weathering environment in dramatically higher resolution than possible by remote analysis of the planetary surface.

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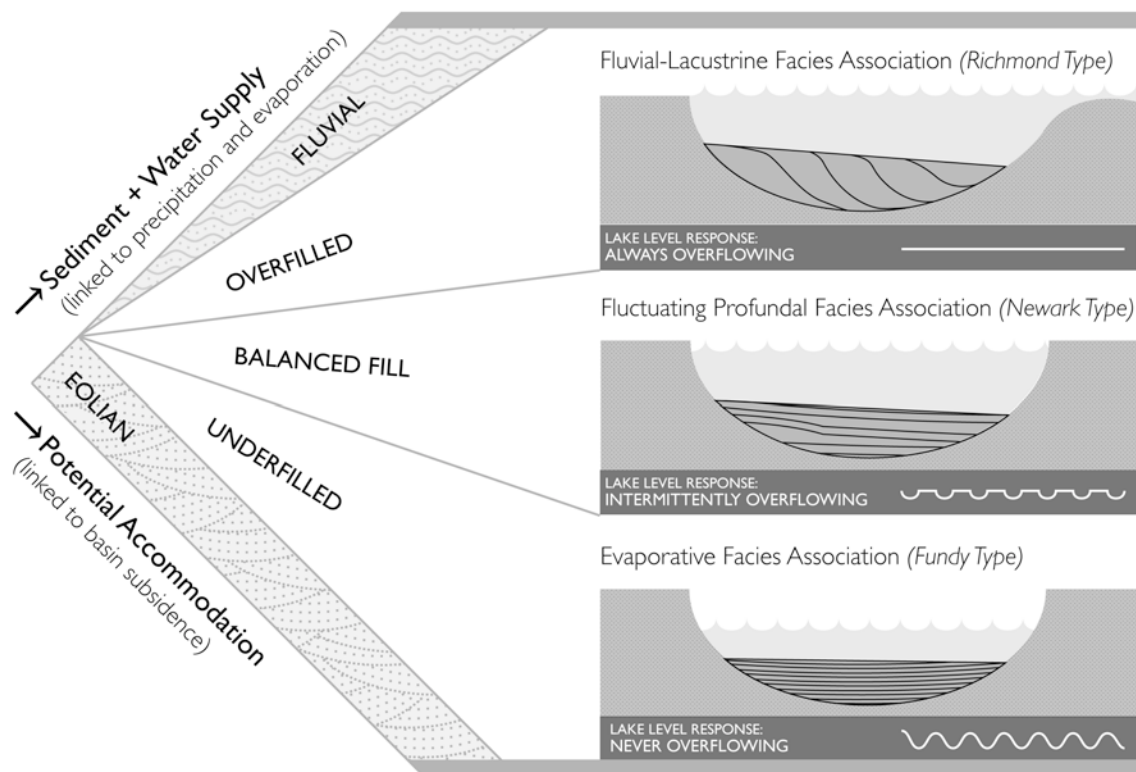


Figure 1: Lake classification system. The lacustrine geologic record is separable into three endmember lithofacies associations that correspond to three basin types: overfilled, balanced fill, and underfilled (e.g., Bradley, 1925; Olsen, 1990; Carroll and Bohacs, 1999). Two primary factors differentiate these lake types: accommodation space and the supply of water and sediment to the basin. This classification scheme provides a powerful framework for analyzing Mars paleolakes by integrating observations of basin structure (e.g., impact craters), outlet-controlled baselevel, and sedimentary deposits. After Carroll and Bohacs (1999).

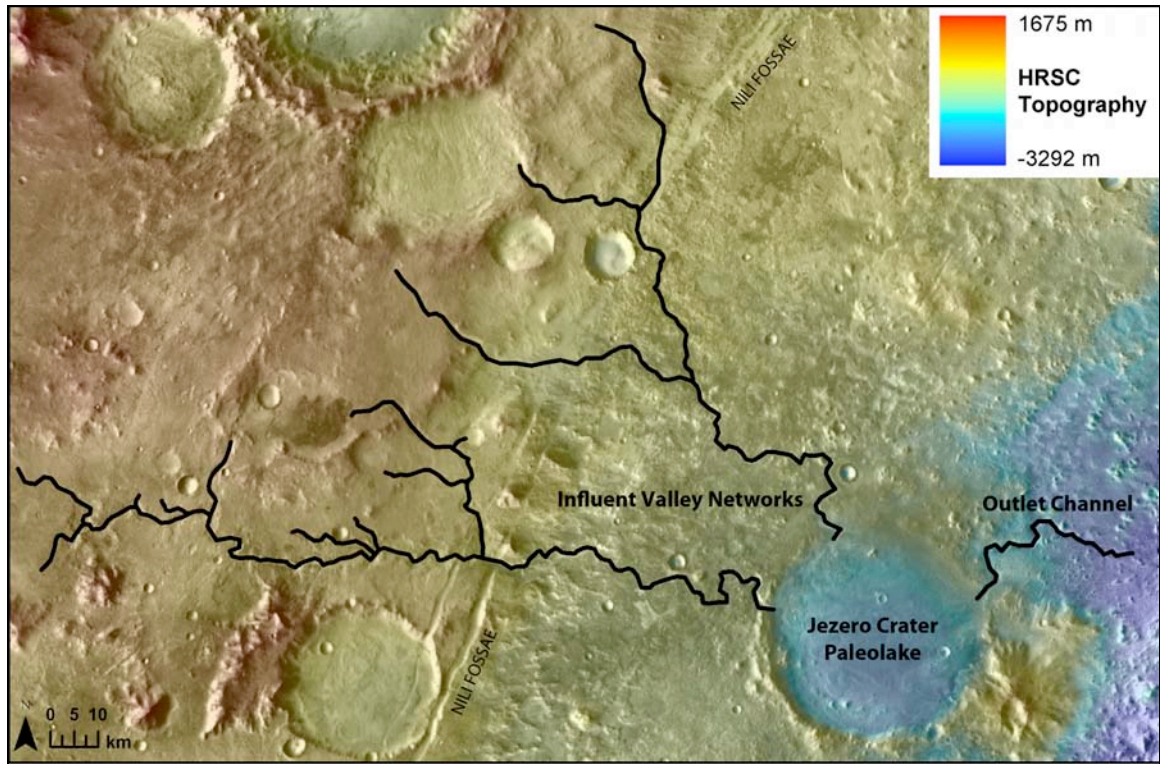


Figure 2: Overview of the Jezero watershed, basin, and outlet channel. The Jezero watershed covers $\sim 15,000\text{-km}^2$ of mineralogically-diverse (e.g., Mangold et al., 2007; Ehlmann et al., 2008a; Ehlmann et al., 2009) Noachian terrane in the Nili Fossae region of Mars (18.4°N , 77.7°E) northwest of Isidis Planitia. Two valley networks with 645 km of channels source the adjoining sedimentary deposit in the 45-km diameter basin (Fassett and Head, 2005). An outlet channel, which controlled baselevel in the lacustrine system, is mapped for ~ 53 kilometers. Topography from HRSC over THEMIS.

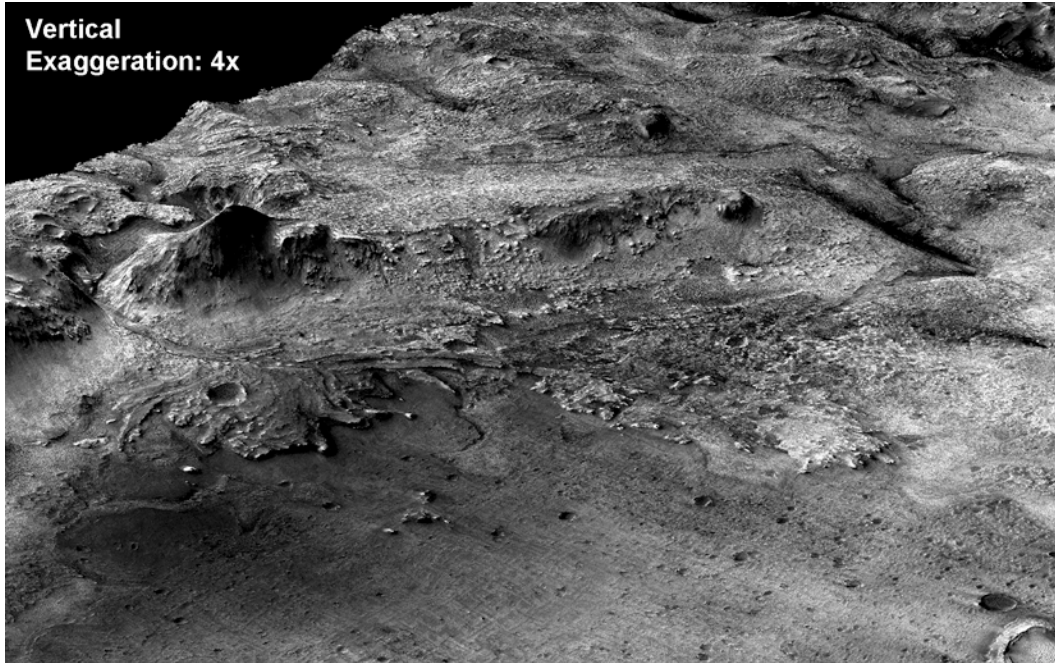


Figure 3: Perspective view of the Jezero sedimentary deposits looking northwest from within the basin. Both valley network inputs are visible at left (west) and right (north). Portions of CTX: P04_002743_1987_XI_18N282W and P03_002387_1987_XI_18N282W over HRSC DTM's from orbits h0988 and h2228 with 4x vertical exaggeration.

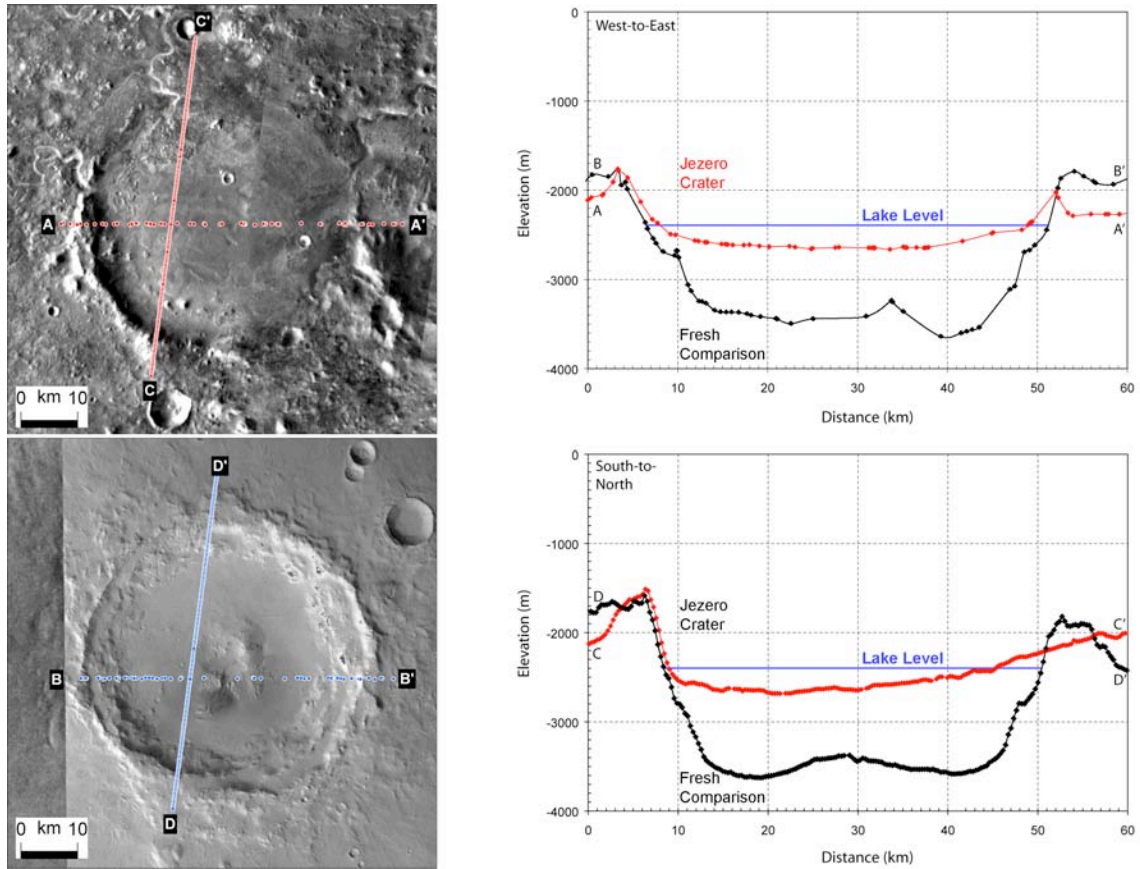


Figure 4: Topographic profiles from MOLA point data of Jezero crater (top) and a similarly sized fresh crater (bottom) found in Elysium Planitia (8.25°N, 125.75°E). This comparison shows that Jezero crater (top) has experienced approximately 1-km of infilling compared to the morphologically fresh crater. Lake level is shown at -2400 m as suggested by the elevation of the outlet.

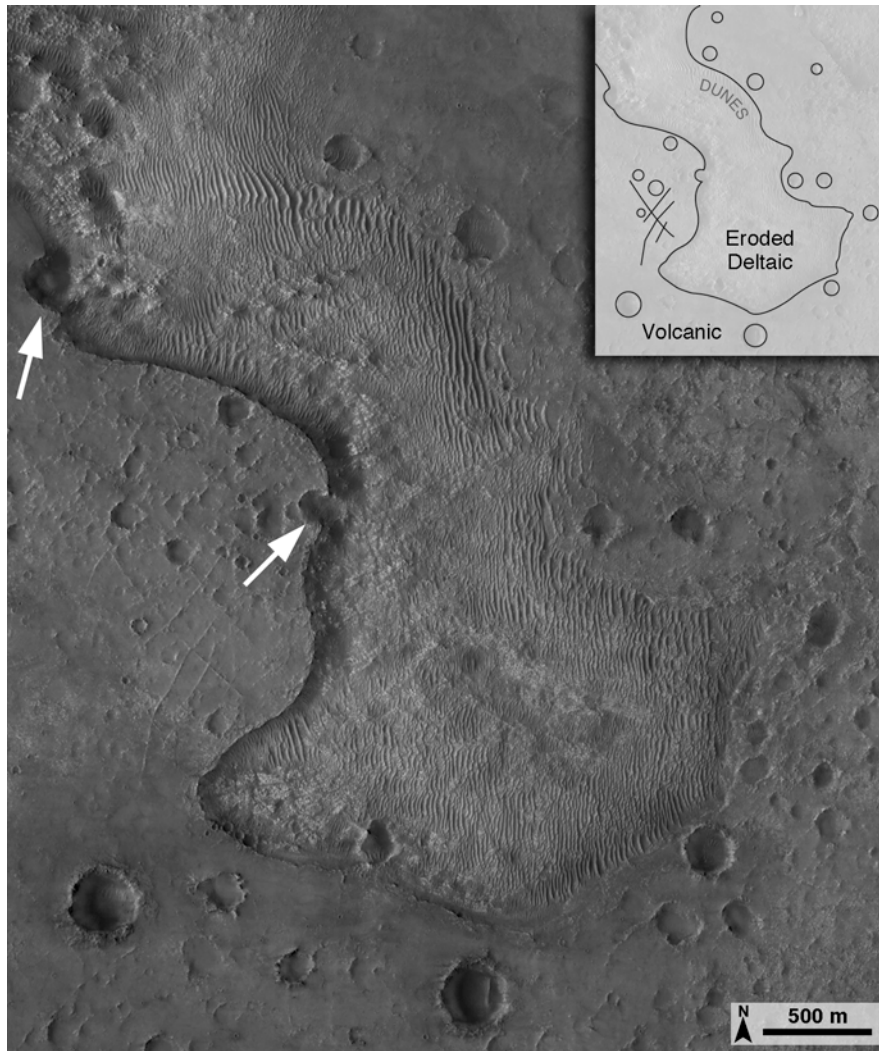


Figure 5: Lava flooding of the Jezero crater floor in areas of eroded deltaic deposits. Directly basinward of the continuous sedimentary deposits in Jezero (Figure 3), embayment relationships indicate that the delta was larger in the past. In this scene, light-toned sedimentary material (with dunes) has been embayed by an early Amazonian (Figure 6) unit interpreted as volcanic. Craters that impacted on the boundary between the competent volcanic material and the weak sedimentary deposit (marked with white arrows) have experienced differential preservation. The previously high-standing deltaic

material eroded to approximately its present configuration prior to formation of the volcanic unit. Portion of HiRISE: PSP_002743_1985.

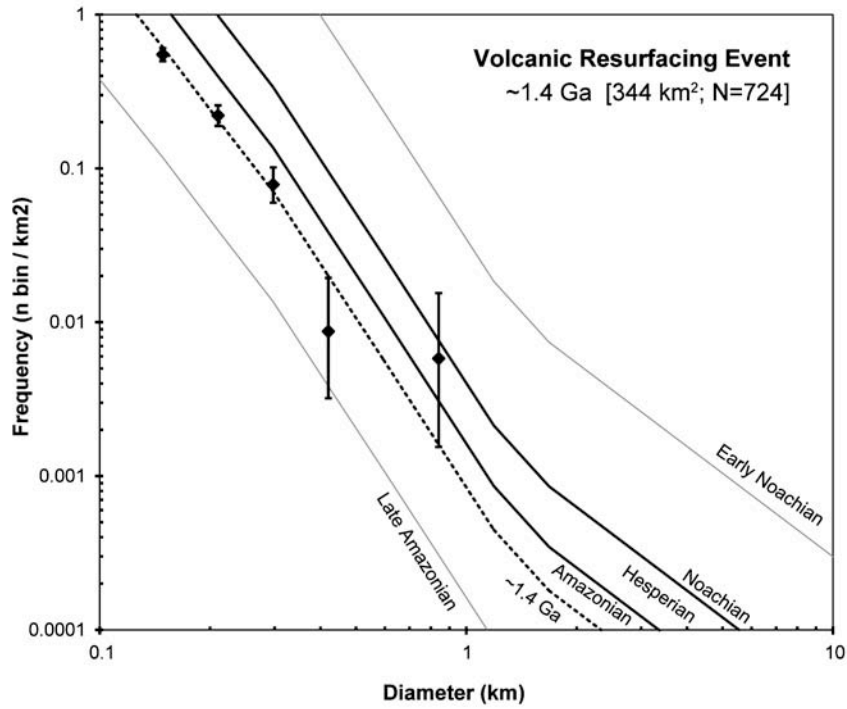


Figure 6: Impact crater-size frequency distribution of craters superposed on the volcanic floor unit (Figure 5). Such resurfacing is not unusual (Goudge et al., 2011). Our crater count reveals 724 craters on 344-km² of this unit. Isochrons of Hartmann (2005) suggest an early Amazonian age (1.4 Ga). This age is consistent with the extensive erosion prior to the resurfacing event as well as the erosion postdating the resurfacing (Figure 5).

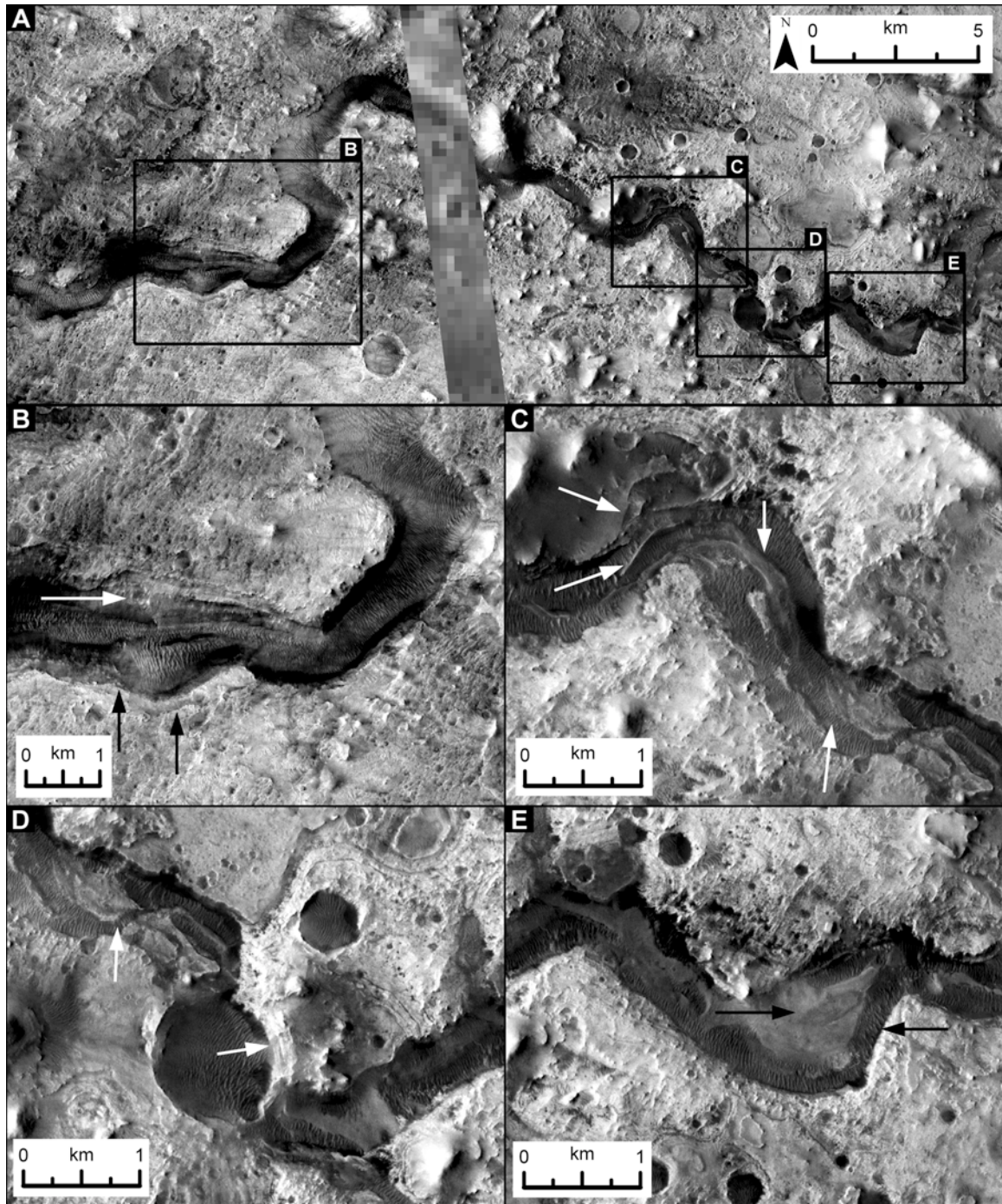


Figure 7: Outlet channel morphology. The outlet channel has a sinuous planform (A) that can be traced for ~53 km eastward from Jezero into a terrane with superposing units (Figure 2). Bar deposits and terraces indicate that this channel did not form from a singular catastrophic breach of Jezero.

The 40-km distance across the basin from the input of the valley networks to the outlet (Figure 2) would have effectively trapped sediment. Therefore, in our interpretation most of the development of sedimentary bedforms in the channel (B-D) is attributable to erosion and deposition by through flow after the initial breaching of the outlet, as shown by arrows. (B) Planar bedding and terraces; (C) Point bar and inner channel; (D) Inner channel and planar bedding exposed by a ~1-km crater; (E) inner channel and massive deposits. Portions of CTX: P15_007068_1971_XN_17N281W and P02_001965_1988_XN_18N281W.

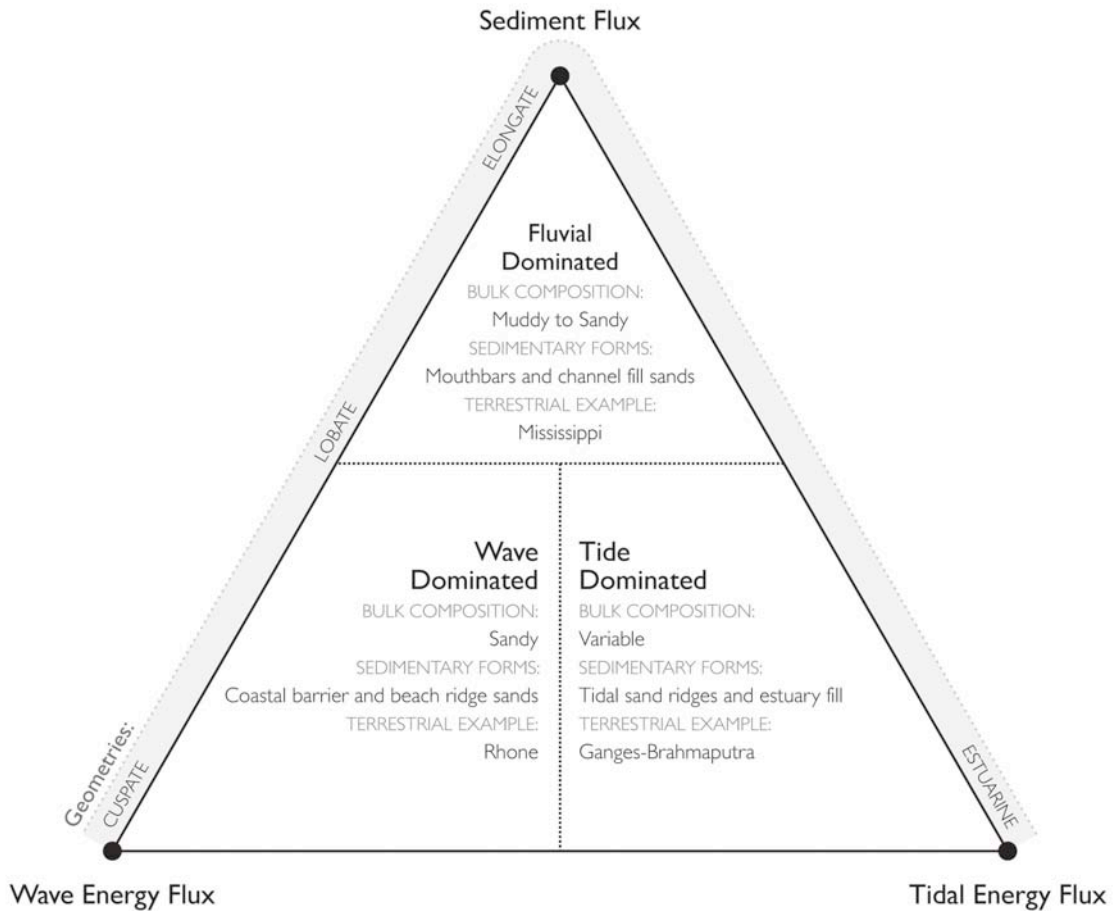


Figure 8: Tripartite classification of deltas. Varying delta morphologies result from the relative influences of the sediment supply, wave energy, and tidal energy (after Galloway, 1975). The Jezero delta is dominated by the sediment supply and has a lobate form (Figure 3). The lacustrine setting minimized the influence of tide and wave energy on the development of the delta.

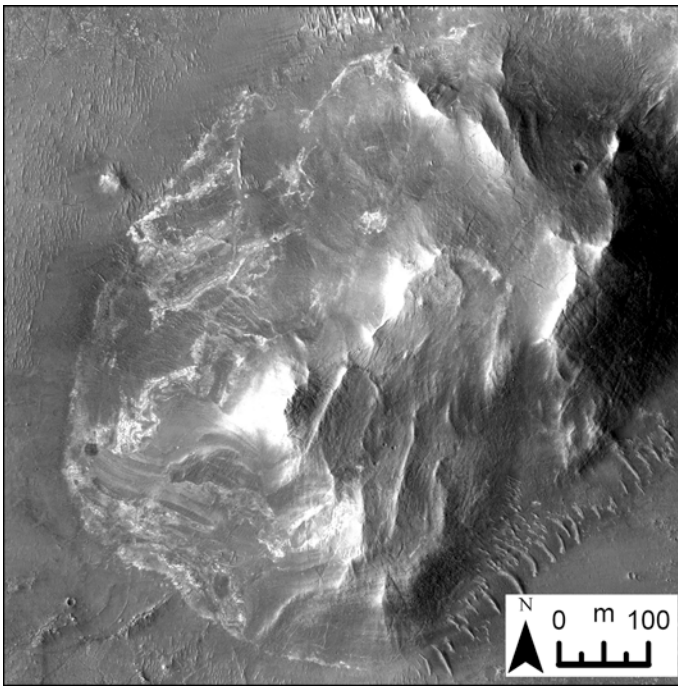
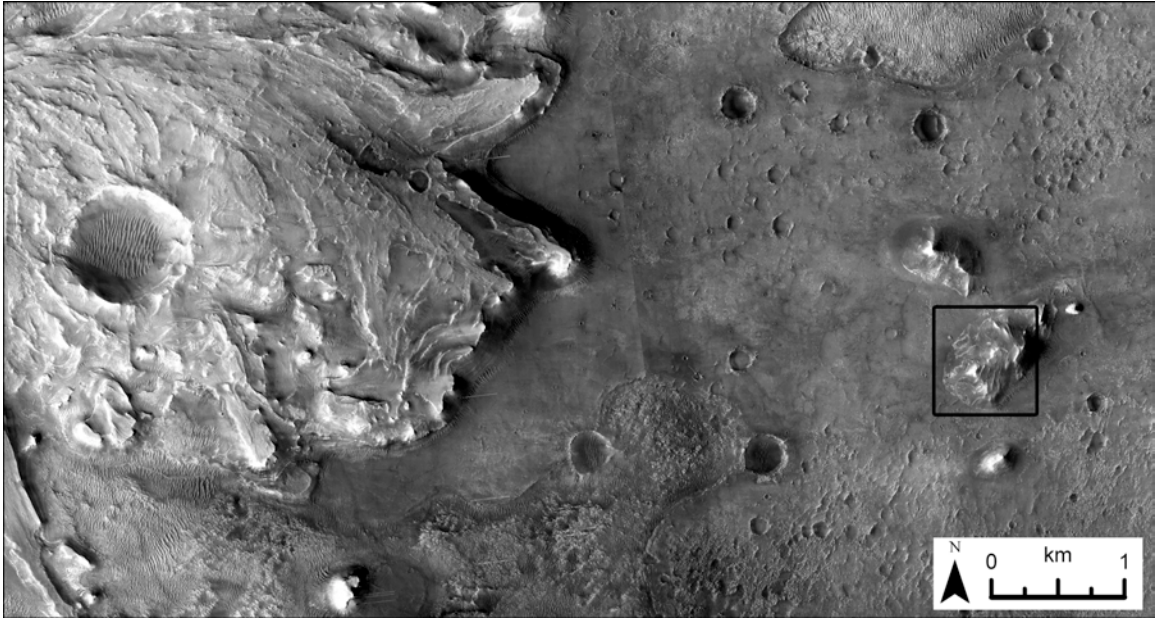


Figure 9: Erosion of the distal part of the delta. (A) Evidence of substantial erosion and resurfacing within the basin is provided by kīpukas of deltaic sediments (at right). These isolated remnants of the delta are entirely surrounded by early Amazonian volcanic material and demonstrate the larger past extent of the delta. The box indicates the area shown in B.

Portions of HiRISE: PSP_002743_1985 and PSP_003798_1985. (B) These remnants of the delta contain layered sediments. The location of these remnants (Figure 9A) and their peak elevation (-2550 m), comparable to the escarpment (-2500 m), require a much larger past extent of the delta. Portion of HiRISE: PSP_002743_1985.

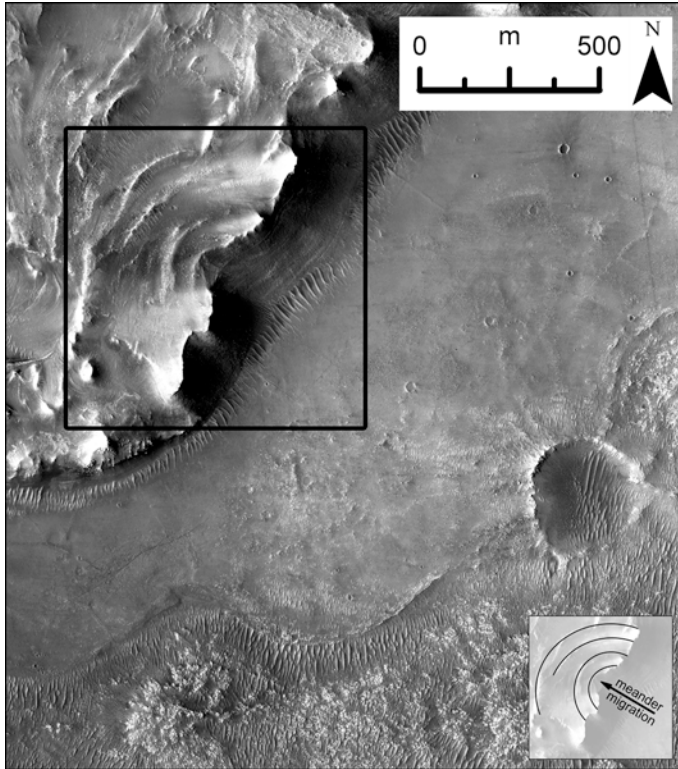


Figure 10: The escarpment of the deltaic deposits truncates a scroll bar. The orientation of this meander indicates that the direction of migration was proximal (toward the watershed, away from the escarpment). This orientation requires a more extensive delta in the past. Portion of HiRISE: PSP_002387_1985.

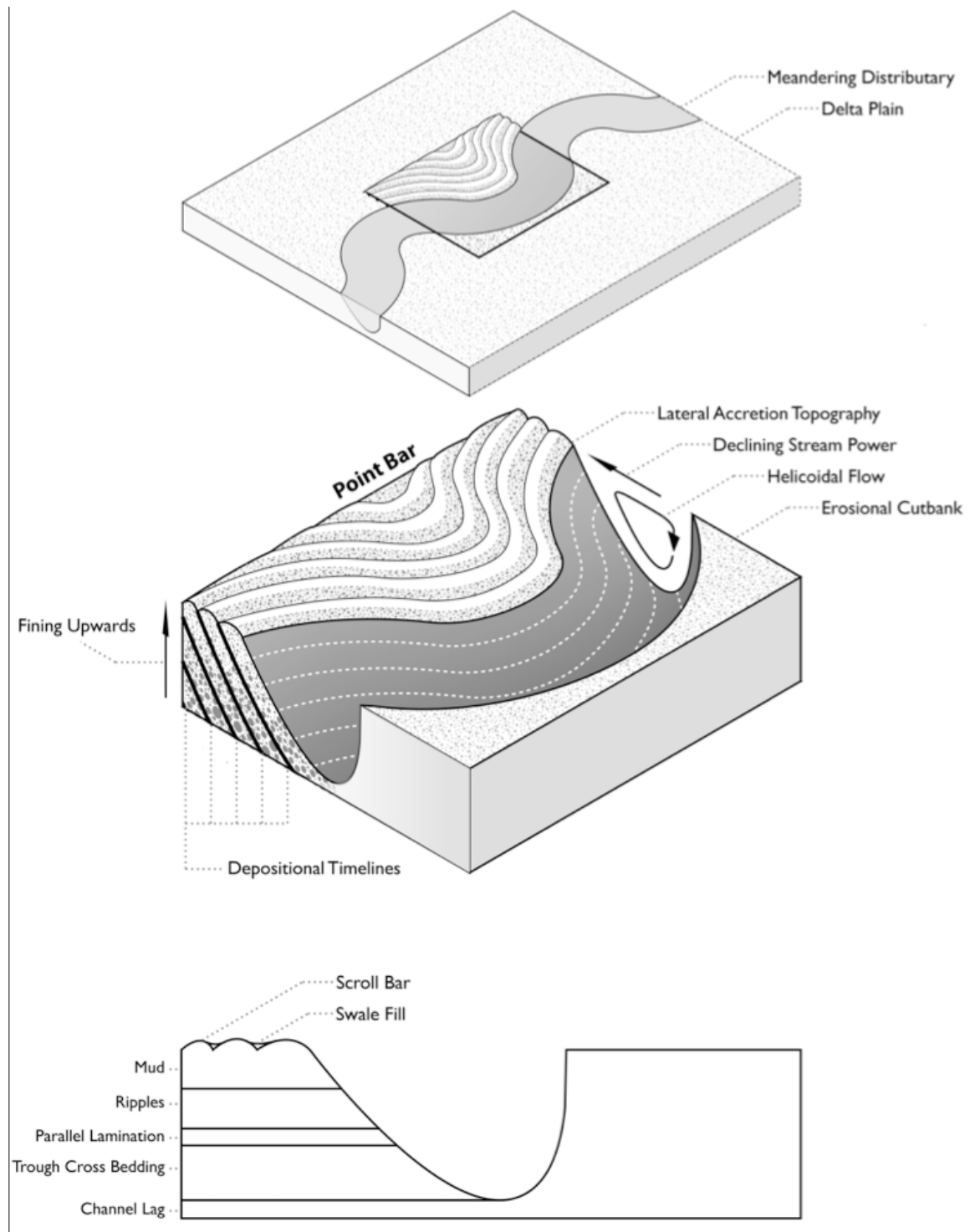
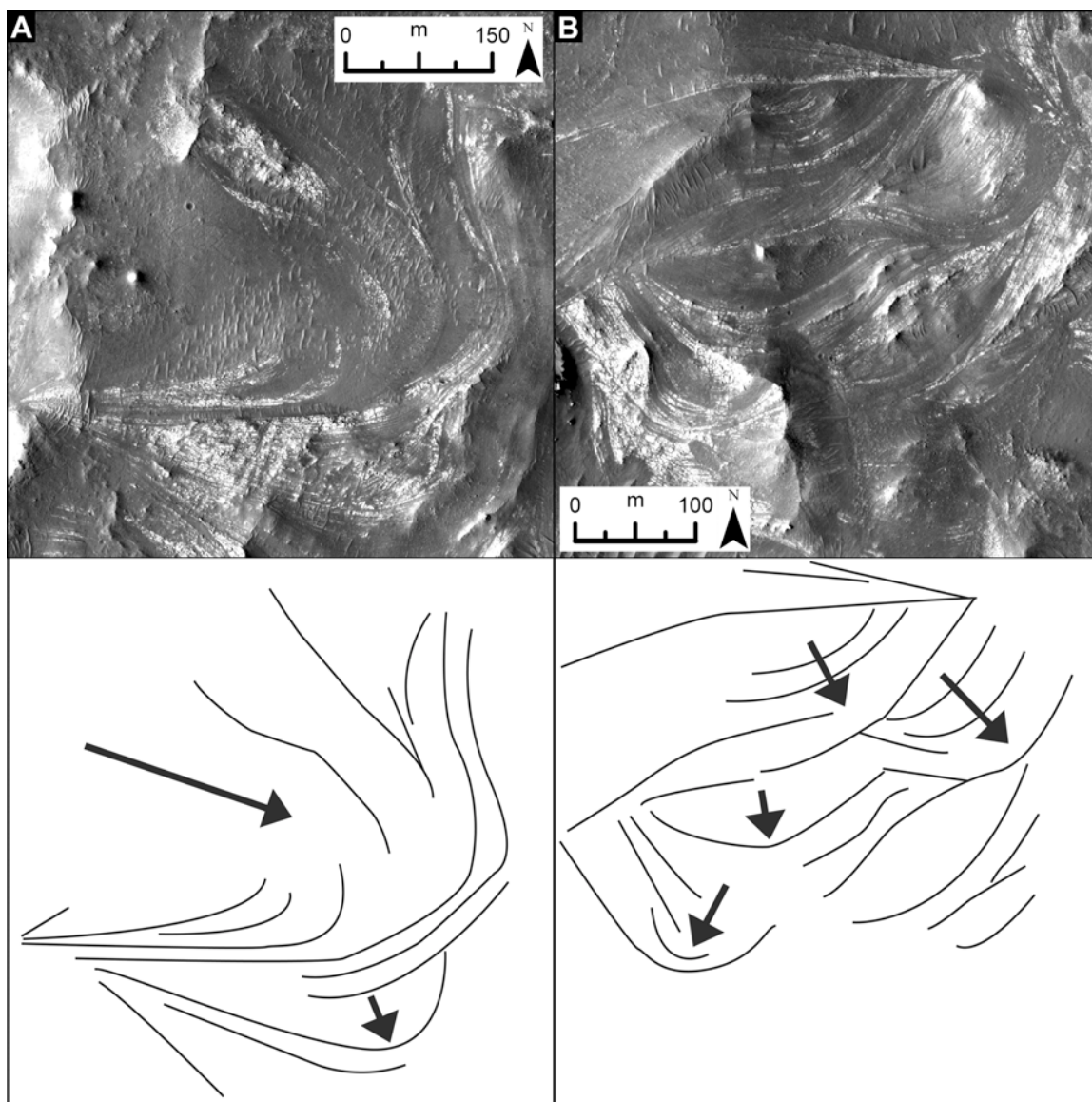
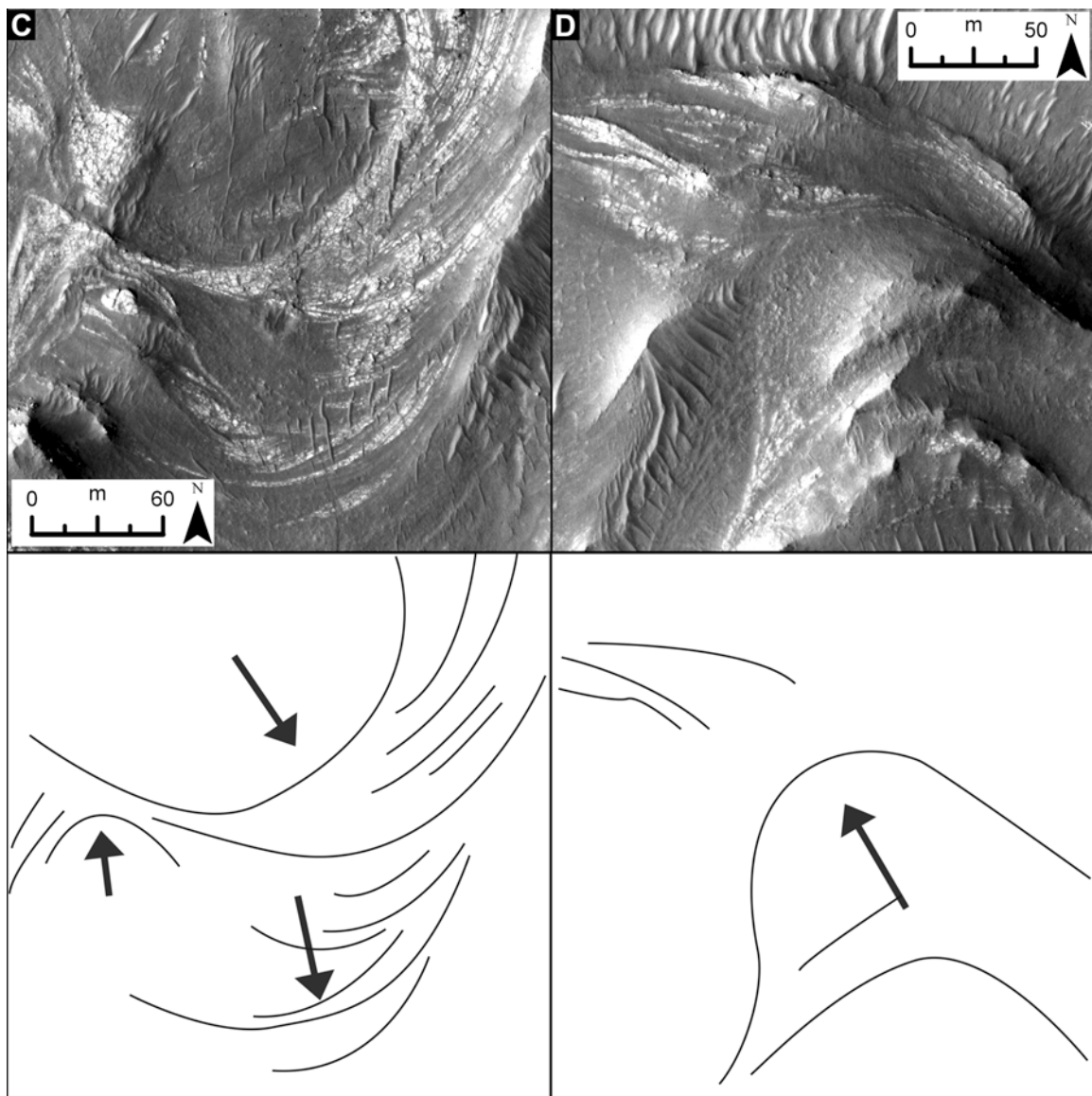


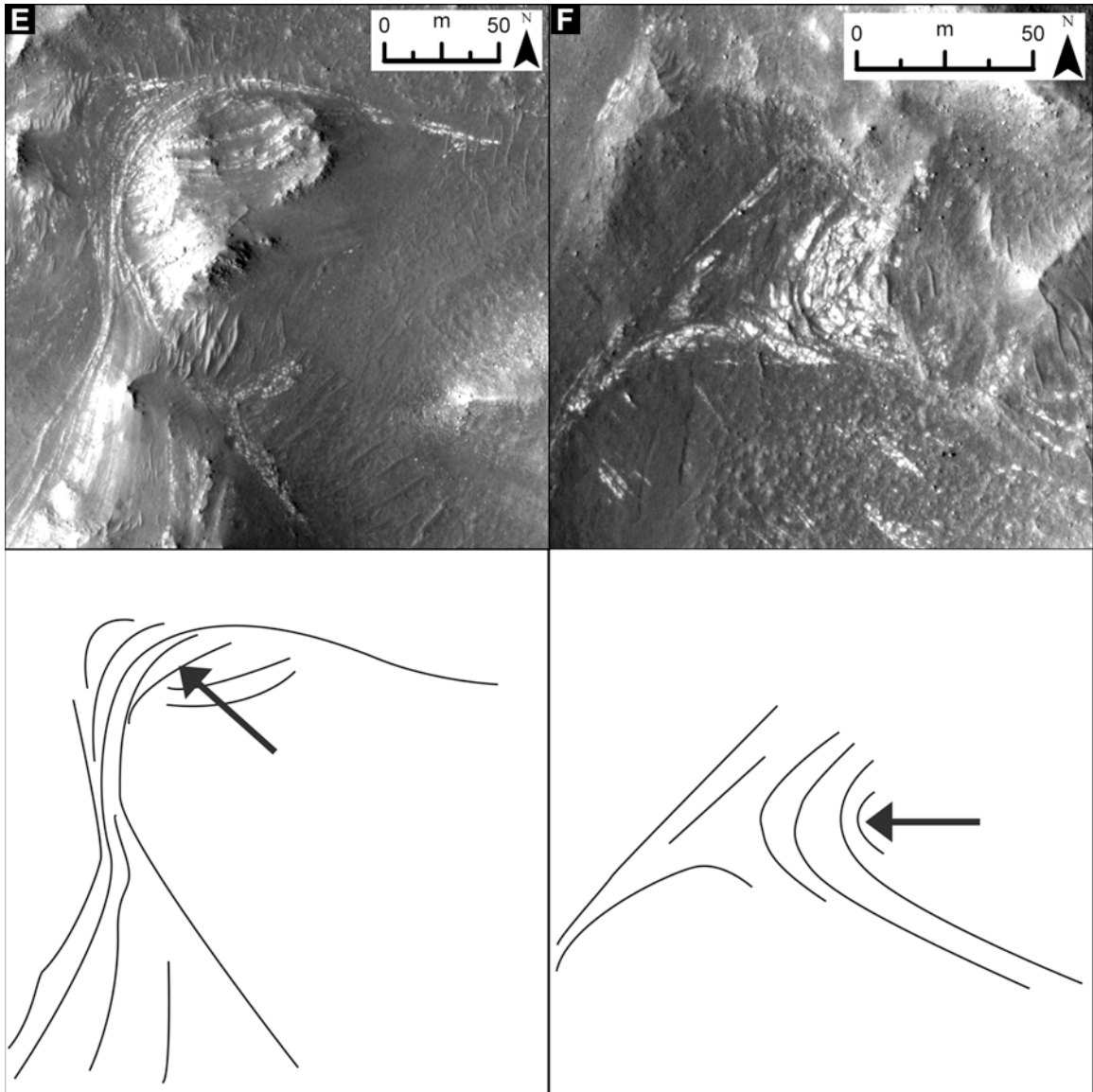
Figure 11: Sedimentary structures and lithologies in meandering streams.

At a meander (top, middle), erosion occurs on the outside of the bend and deposition of a point bar sequence occurs on the inside of the bend, forming

lateral accretion topography. Scroll bars (bottom) are the characteristic planimetric signature of lateral accretion. Point bars are prograding, diachronous, time-transgressive, laterally continuous, fining upward sequences. Planar lateral accretion surfaces (depositional timelines) within the point bar (middle) dip in the direction of channel migration and are responsible for epsilon cross-bedding (Figure 13). Current direction is parallel to the strike of lateral accretion surfaces.







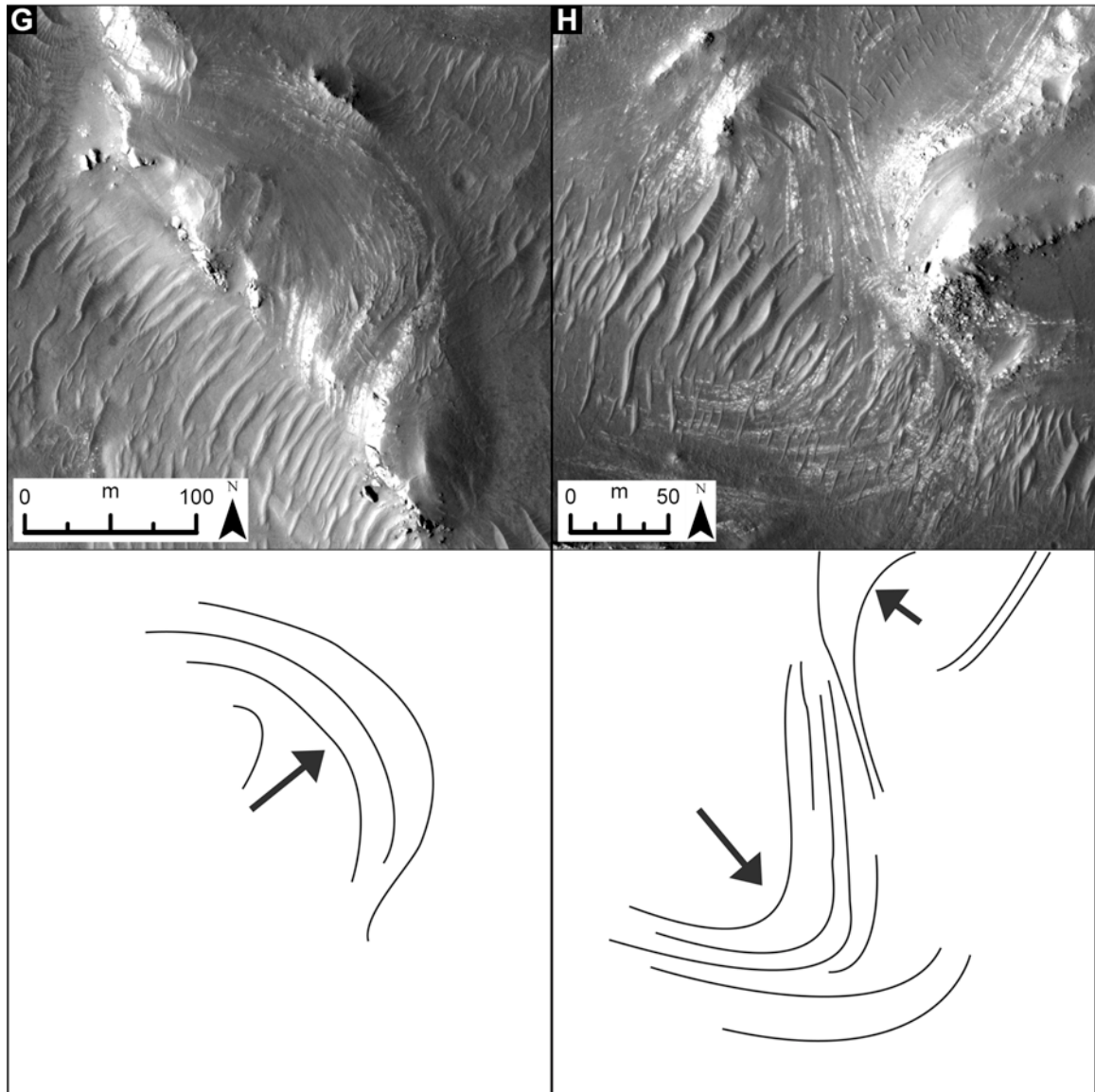
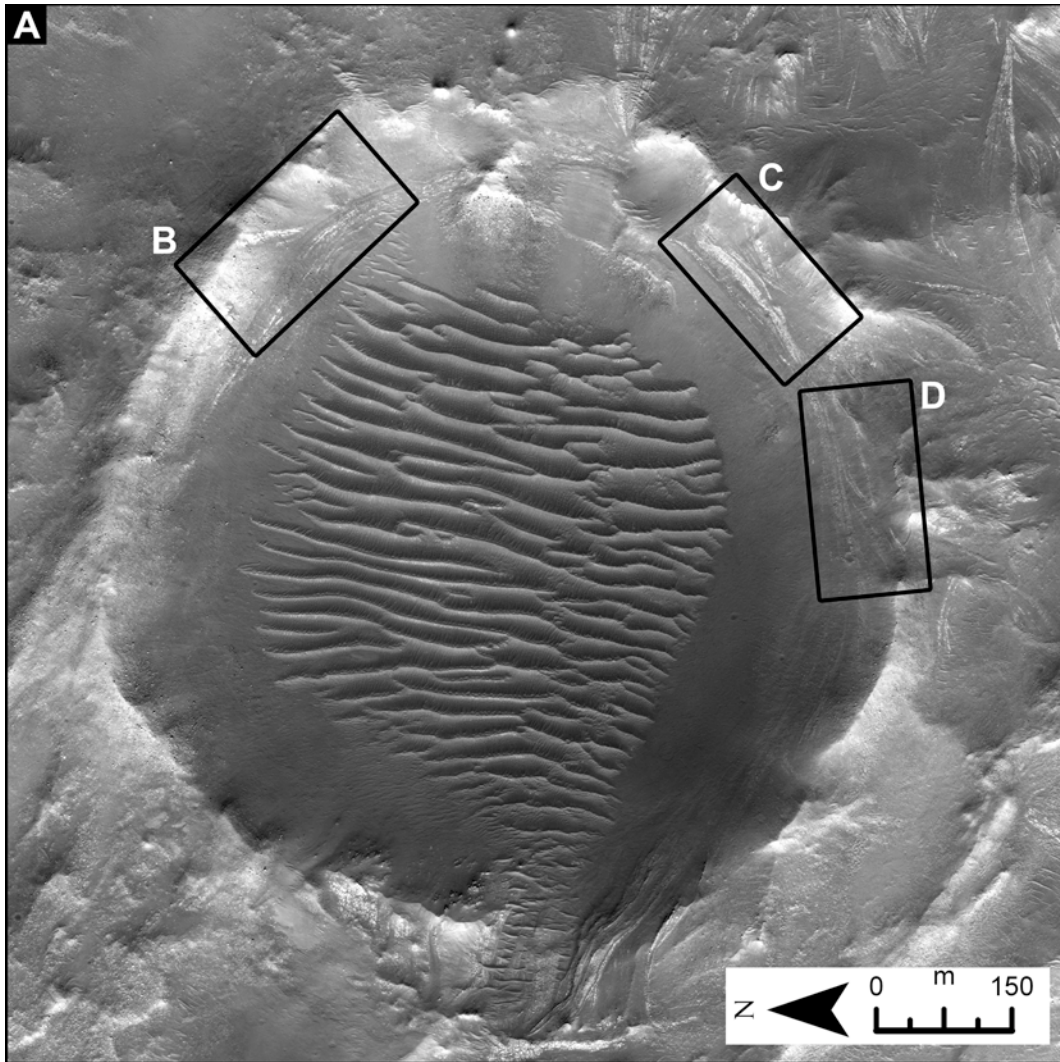
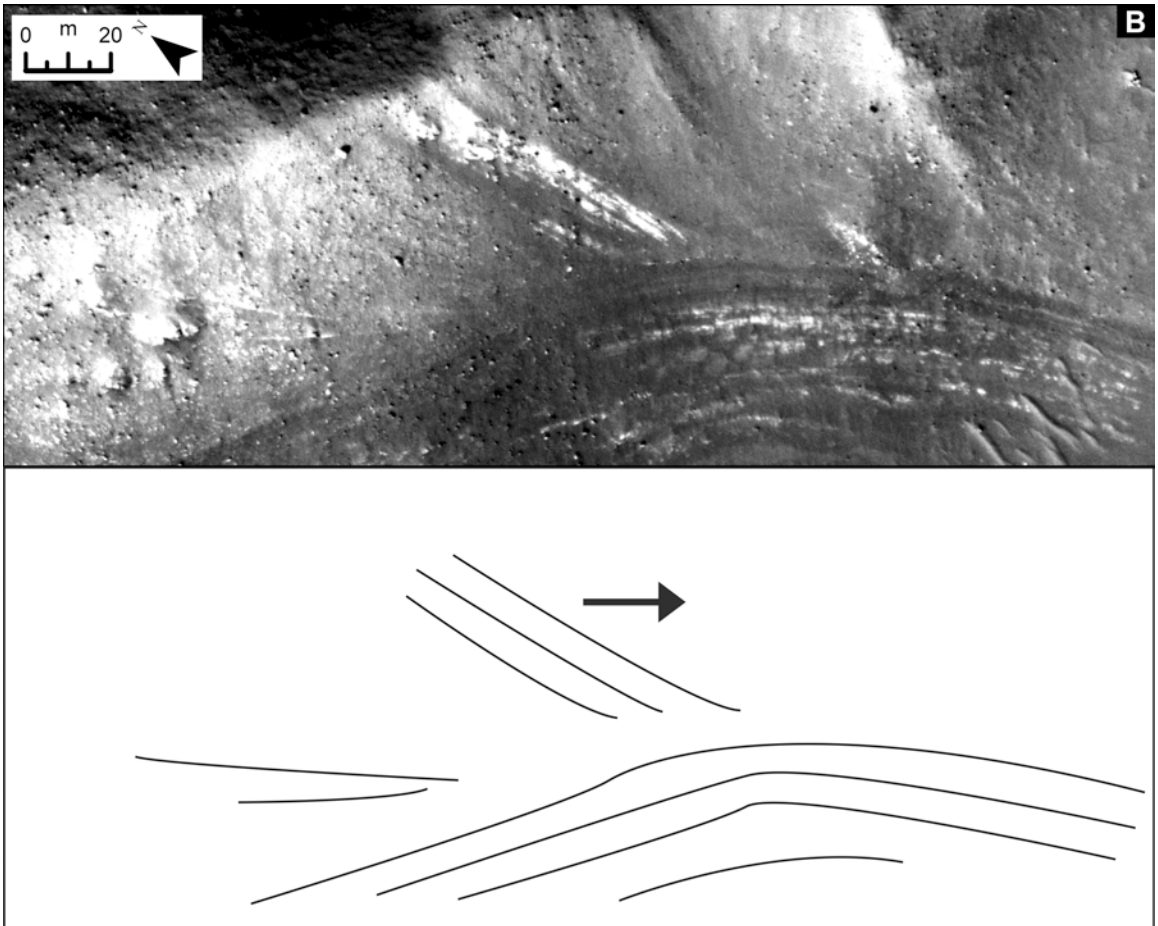
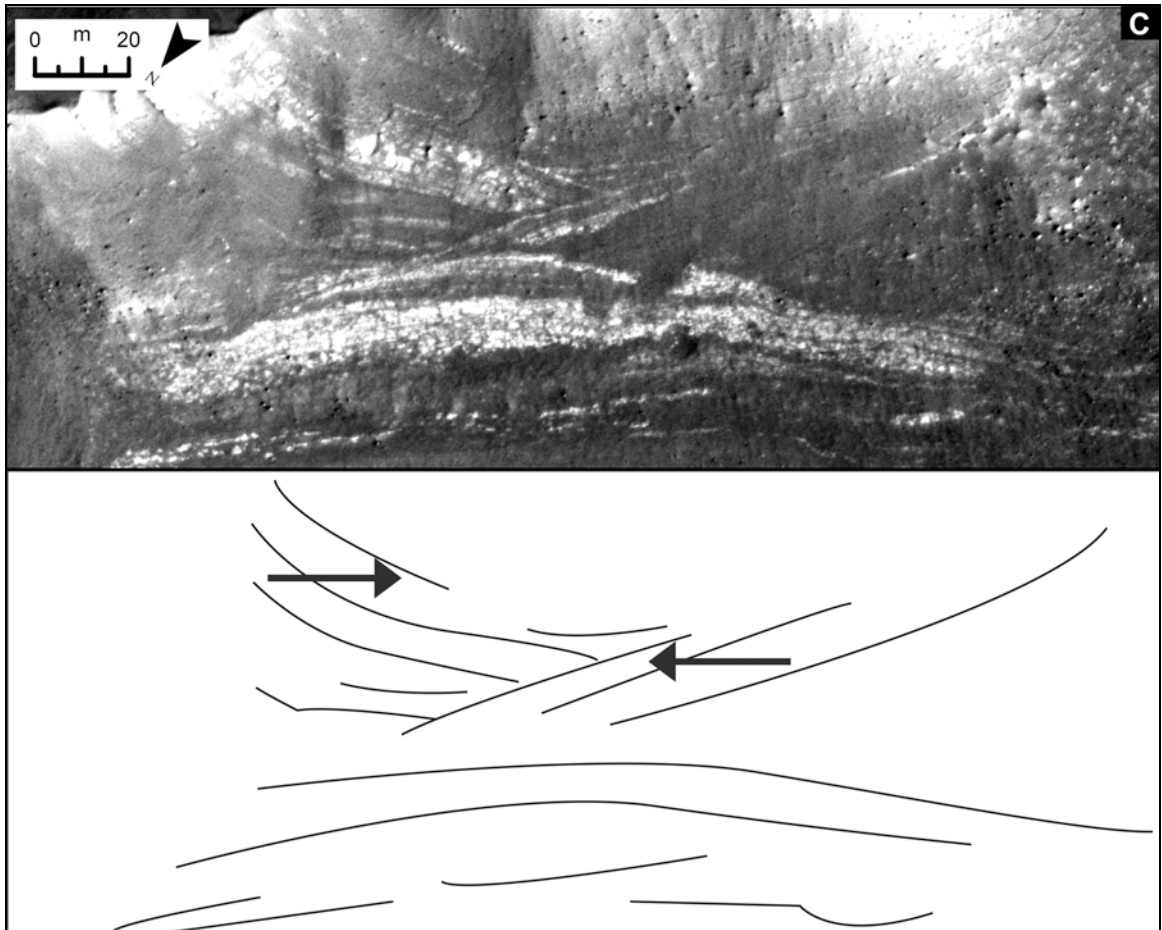


Figure 12: Scroll bars. Numerous scroll bars (lateral accretion topography, Figure 11) are observed adjacent to the one-km crater on the delta plain (Figure 13) and stratigraphically below erosionally-resistant materials interpreted as channel sands (Figure 14). These features (A-H) result from the development of point bars at the inside of distributary channel meander bends (Figure 11). Arrows indicate the direction of channel migration.

Unconformities in the scroll bar patterns are common and suggest successive channel migrations. Portions of HiRISE: PSP_002387_1985.







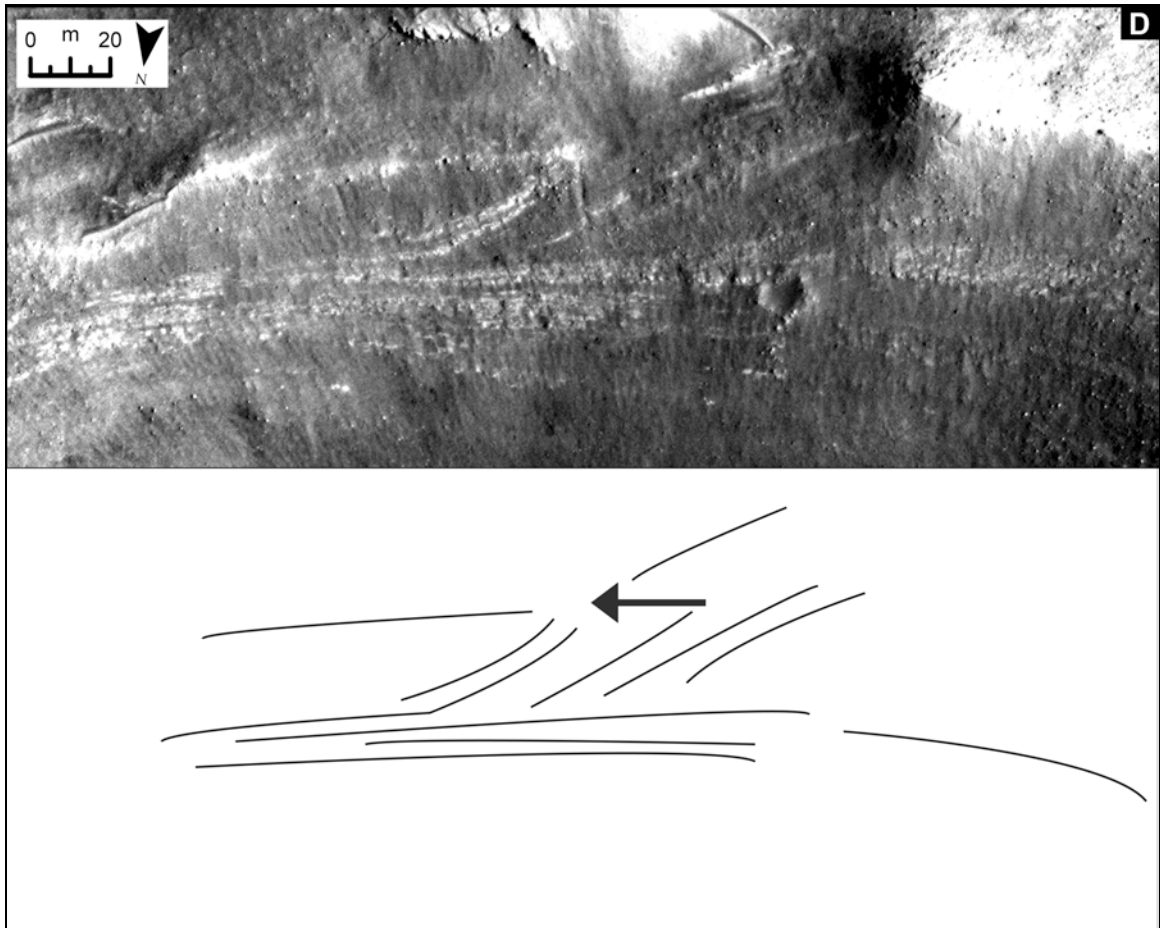


Figure 13: Deltaic deposit cross-sections. (A) A one-kilometer diameter crater provides a cross-sectional view of the deltaic sedimentary materials. The crater is oriented in this view such that north is to the left and basinward toward the top. In the walls of this crater, epsilon cross-bedding (Allen, 1963) is observed. Epsilon cross-bedding results from lateral accretion surfaces within point bars (Figure 11, Figure 12). Portion of HiRISE: PSP_002387_1985. (B) Within the one-kilometer diameter crater, three clear examples of epsilon cross-bedding are observed (A,B,C). In these outcrops lateral accretion is observed in both the basinward direction (A, C) as well as laterally (B), consistent with our interpretations of meandering distributaries

of a subaerial delta plain environment. Lateral accretion surfaces dip in the direction of channel migration (indicated by arrows). The paleocurrent direction is parallel to the strike of the lateral accretion surfaces. Portion of HiRISE: PSP_002387_1985.

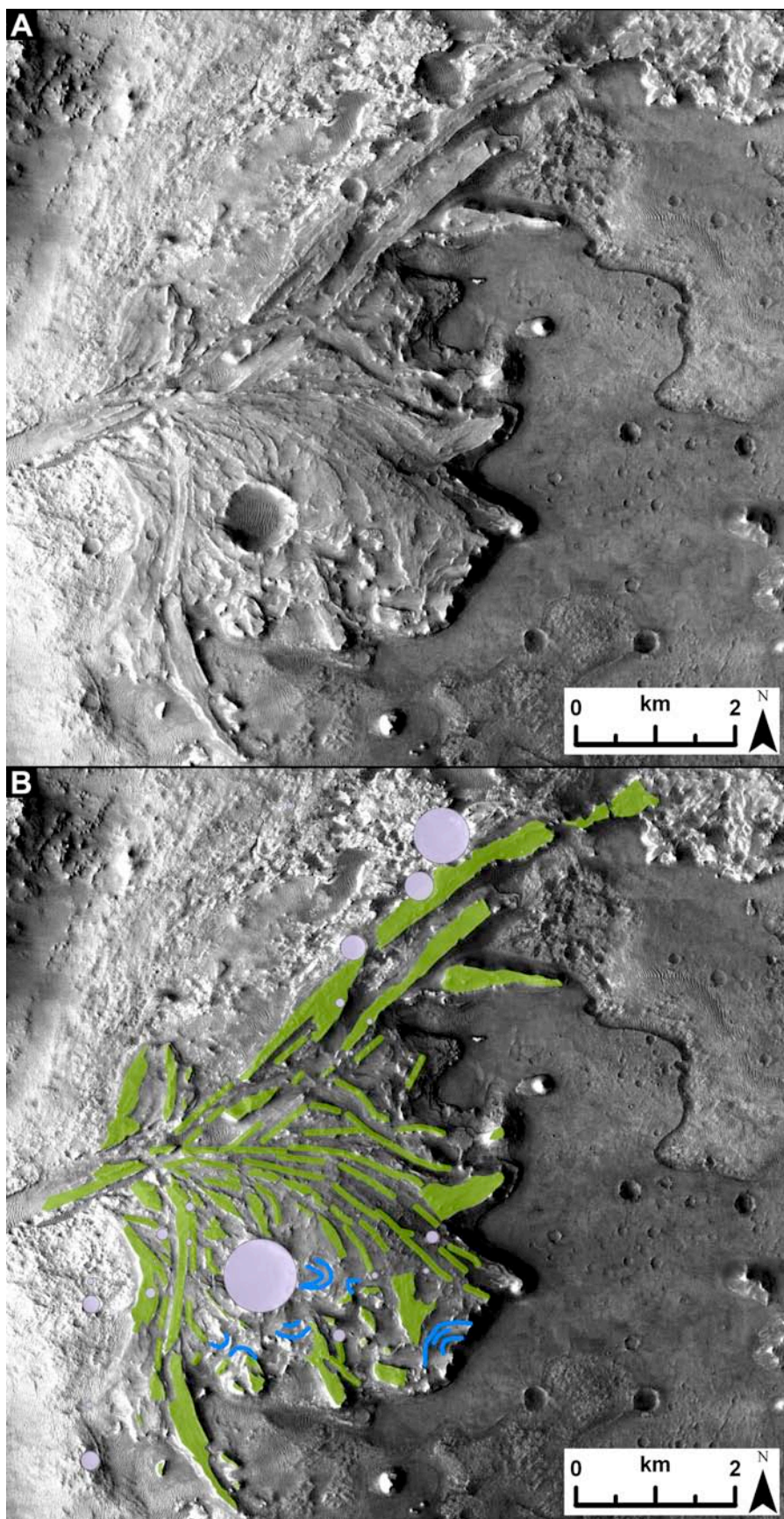
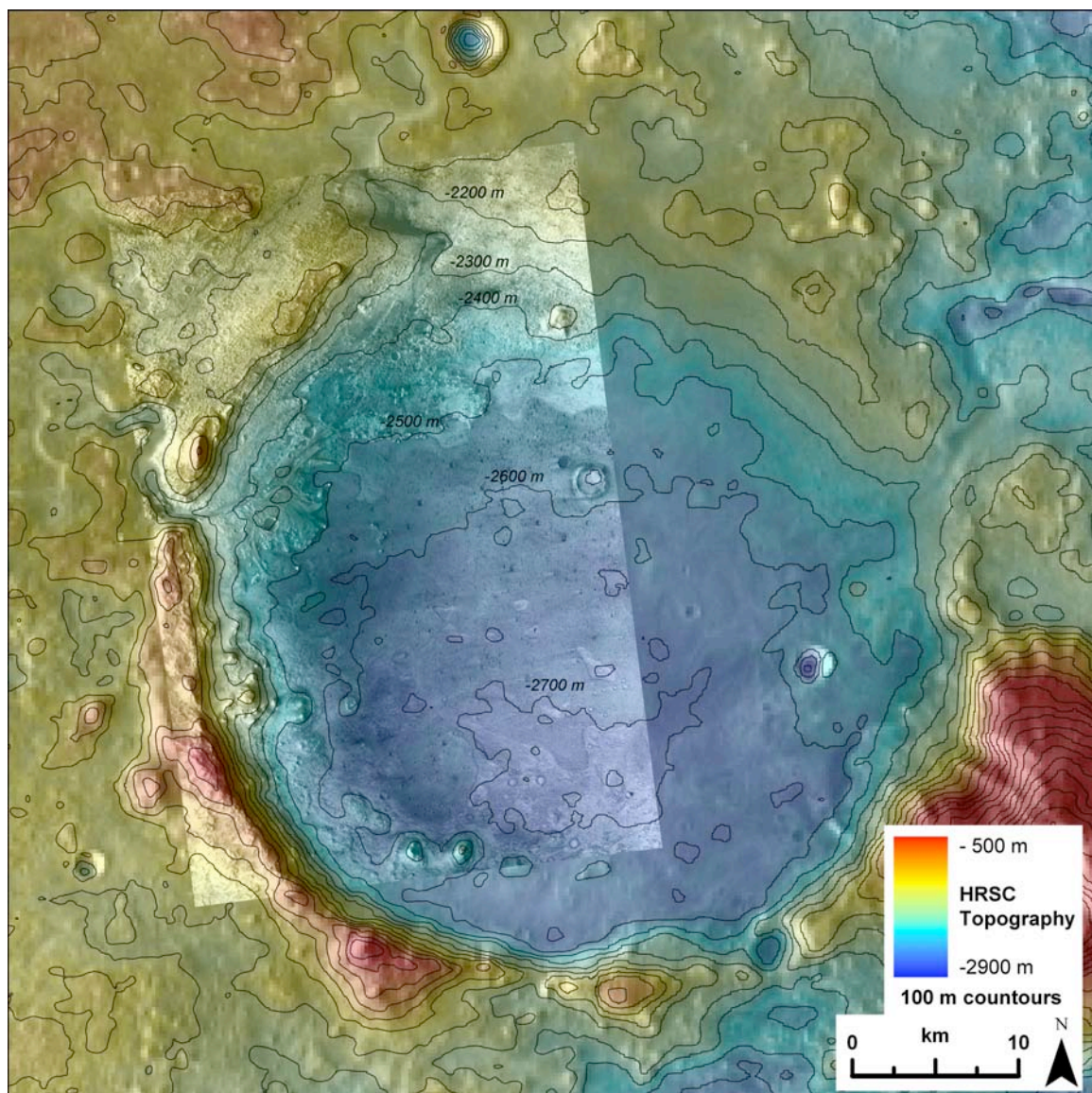


Figure 14: Channel deposits. (A) Stratigraphically above the scroll bars (Figure 12) and epsilon cross-bedding (Figure 13) are elongate erosionally resistant materials that we interpret as channel sands. Consistent with our interpretation of a more extensive delta in the past (e.g., Figure 10), these channel sands would have been deposits in distributary channels sourcing more distal depocenters. In our interpretation these channel sand deposits are high-standing because they are more erosionally resistant than overbank deposits. (B) Channels sands are mapped in green, scroll bars in blue, and craters in purple. Portion of CTX: P04_002743_1987_XI_18N282W.



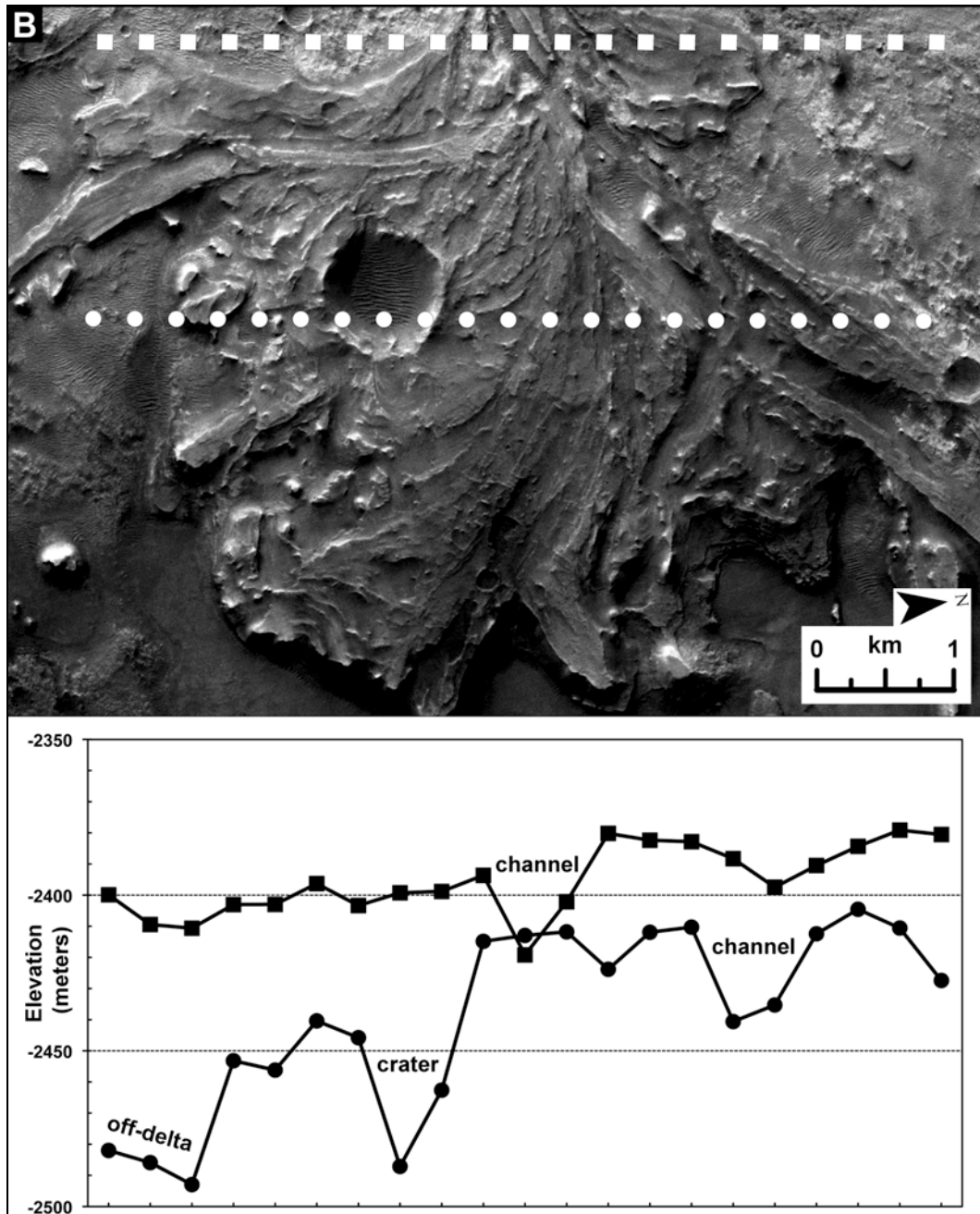


Figure 15: (A) Topographic map with 100 m contours from HRSC. (B) Topographic profiles across the delta reveal the stratigraphic position of the channel sands (Figure 14) above the scroll bars (Figure 12) and epsilon cross-bedding (Figure 13). Portion of CTX: P04_002743_1987_XI_18N282W. MOLA Orbits: 15454 and 15127.

CHAPTER 2

UNIQUE CHRONOSTRATIGRAPHIC MARKER IN DEPOSITIONAL FAN STRATIGRAPHY ON MARS: EVIDENCE FOR ~1.25 Ma GULLY ACTIVITY AND SURFICIAL MELTWATER ORIGIN.

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chronostratigraphic marker in depositional fan stratigraphy on Mars:
Evidence for ca. 1.25 Ma gully activity and surficial meltwater origin.

Geology, 37, 207-210, doi: 10.1130/G25398A.1.)

Abstract

The origin of gullies on Mars is controversial (e.g., catastrophic groundwater release, debris flows, dry granular flows, or meltwater from surface ice and snow) and their ages are difficult to determine due to their small size. We describe a gully depositional fan that contains a unique chronostratigraphic marker (secondary crater clusters) between episodes of gully activity during fan development. This marker can be traced to its source, a 7-km-diameter rayed crater that we have dated as ca. 1.25 Ma. This age links gully activity to the emplacement of dust-ice mantling deposits interpreted to represent recent ice ages on Mars. This association, together with multiple episodes of depositional fan formation, favors an origin for

these gullies from top-down melting of snow and ice during multiple favorable spin-axis and orbital variations. This melting mechanism is consistent with the occurrence of gullies in unique steep-sloped, poleward-facing insolation microenvironments that favor the melting of small amounts of surficial snow and ice.

Introduction

Martian gullies were defined by Malin and Edgett (2000) as geomorphic features having alcoves that taper into channels that continue downslope to triangular aprons of deposited material. They interpreted these features as consistent with fluvial erosion and proposed that groundwater release from confined aquifers was responsible for gully formation, a hypothesis further developed by Heldmann and Mellon (2004) and Heldmann et al. (2007). Many studies subsequent to Malin and Edgett (2000) have identified additional examples of gully morphology and have further characterized the geographic distribution of gullies in the middle and high latitudes of both hemispheres and their geologic settings on crater and valley walls, mesas, central peaks, and exterior crater rims (e.g., Balme et al., 2006; Dickson et al., 2007).

Since their discovery, a variety of formation hypotheses has been proposed to explain the diversity of gully observations. These hypotheses can be divided into three broad categories: entirely dry mechanisms (e.g.,

Treiman, 2003; Shinbrot et al., 2004), wet mechanisms invoking groundwater release (e.g., Malin and Edgett, 2000; Mellon and Phillips, 2001; Heldmann and Mellon, 2004), and wet mechanisms invoking surficial meltwater (e.g., Costard et al., 2002; Christensen, 2003; Head et al., 2008). It has been difficult to differentiate between these hypotheses and test their validity using past observations (e.g., Pelletier et al., 2008). Also uncertain is the age of Mars gullies and thus their specific link to recent climate history. Although they appear to have formed contemporaneously with latitude-dependent mantling deposits thought to have been emplaced during recent “ice ages” (e.g., Mustard et al., 2001; Head et al., 2003; Milliken et al., 2003; Reiss et al., 2004; Kostama et al., 2006), the area of individual gullies is too small to obtain reliable ages using crater size- frequency distributions.

In this study we document evidence of a gully system (Fig. 1) that contains secondary craters within its depositional fan that can be traced back to the primary crater, which can be reliably dated, thereby providing a chronostratigraphic marker for an intermediate stage of gully development. Stratigraphic relationships in the depositional fan of this gully system suggest (1) multiple episodes of alluvial fan-style deposition, (2) very recent depositional activity that is younger than a newly recognized rayed crater that is the source of the secondary craters, (3) temporal links to the recent climate history, and (4) surficial snowmelt as the most likely source of these multiple episodes of recent gully activity.

Gully Fan Stratigraphy

In eastern Promethei Terra ($\sim 35^{\circ}\text{S}$, 131°E), there is an $\sim 5\text{-km}$ -diameter crater with a single well-developed gully system (Fig. 1) and several smaller gullies in its north-northeast wall. The smaller gullies lack discernible alcoves (i.e., they are channel fan assemblages), their incision is limited to the surficial mantling deposit (Mustard et al., 2001; Head et al., 2003; Milliken et al., 2003), and their fans clearly superpose the crater-fill floor material. In contrast, the large gully system (composed of a small western gully and larger eastern gully) shows evidence for incision into the crater wall and has multiple contributory subalcoves and channels (Fig. 1).

The low-slope depositional fan associated with this gully system is significantly larger than the others and is bounded on its western margin by a small arcuate ridge. Depressions with similar bounding ridges are commonly observed in this latitude band ($\sim 30\text{--}50^{\circ}\text{S}$) in association with deeply incised gully alcoves that are interpreted as the accumulation zones for cirque-like glacial systems (Head et al., 2008). Therefore, the ridge is likely a moraine-like structure bounding a glacially-formed depression into which the fan is deposited, analogous to similar stratigraphic relations described by Hartmann et al. (2003), Berman et al. (2005), and Head et al. (2008).

The gully fan is composed of multiple lobes with distinct lobe contacts, incised channels, and channel fill deposits, all features similar to those observed in terrestrial alluvial fans, i.e., cone-shaped deposits of fluvially transported sediments that accumulate at distinct breaks in slope (Blissenbach, 1954; Blair and McPherson, 1994). Secondary craters (~1–25 m diameter) are pervasive in the vicinity of the gully, but only a portion of the fan has superposed secondaries, implying that at least some portions of the depositional fan were deposited both before and after the emplacement of the secondaries. The individual depositional lobes of the fan can be divided into two groups (Fig. 2): a lobe that predates the secondary crater population (1) and younger lobes (2–4), distinguished by stratigraphic contacts and crosscutting relationships, that are superposed on the lobe with secondary craters. These multiple lobes that postdate the secondary crater population make the emplacement date of the secondary craters a robust maximum age for the youngest lobes of this fan, and therefore the most recent activity of the gully system. We now explore the source of the secondary craters to assess the age of this chronostratigraphic marker in the history of gully activity.

Nearby Rayed Crater Source of Secondary Craters

Regional reconnaissance was undertaken to determine the origin of the secondary craters utilizing the orientation of the crater clusters and cluster patterns. This search led to the discovery of a previously unidentified rayed

crater complex (Fig. 3A) consisting of two superposed very fresh craters, an ~18-km-diameter outer crater and an ~7-km-diameter inner crater located at ~35°7'S, 129°4'E, ~100 km southwest of the gully system. Distinctive rays are observed in THEMIS (Thermal Emission Imaging System) nighttime thermal infrared data (Fig. 3A), but are not observable as albedo contrasts in visible data, consistent with other identifications of young rayed craters on Mars (McEwen et al., 2005; Tornabene et al., 2006). This crater complex is also located in a similar thermophysical setting to previously identified rayed craters, i.e., intermediate albedo and thermal inertia, implying an intermediate dust cover (Mellon et al., 2000).

The morphology of these two impact craters that are candidates for the source of the secondary craters at the gully site reveals their relative states of degradation and modification, and thus which is the most likely candidate for the gully-related secondaries. Both the outer and inner craters have classically defined gullies, preferentially developed within their pole-facing walls. The inner crater gullies have branched subalcoves with intervening sharp spur-like ridges. The channel and alcove walls are smooth and are predominately unconsolidated sediment except where distinct bedrock outcrops occur. The equator-facing wall is a uniform slope of fine-grained material that has formed landslide deposits in several locations and is steeper (~29° versus ~20°) than the gully fans. Neither the inner crater gullies of the pole-facing wall nor the equator-facing wall are blanketed by

latitude-dependent mantling deposits, suggesting that this crater is younger than the most recent episode of latitude-dependent ice-rich mantle deposition at this latitude (Head et al., 2003). The inner crater floor contains a small collection of dunes and landslide materials, but no sublimation features are observed that would indicate glacial modification (Kreslavsky and Head, 2006).

In contrast, morphological observations of the outer crater suggest that it predates deposition of latitude-dependent mantling deposits that would obscure fresh crater rays. Mantling deposits are observed on the rim and walls of the crater, while polygonally patterned ground, indicative of an icy substrate and extended thermal cycling, is observed in gully alcoves (Mangold, 2005; Levy et al., 2009). Furthermore, the outer crater also has a lower depth:diameter ratio (0.07) than the inner crater (0.12), and the walls of the outer crater are of lower slope and have greater asymmetry compared to the inner crater, indicative of more extensive modification (Garvin et al., 2003; Kreslavsky and Head, 2006). Mapping of the rays also shows that their radial distribution is focused within the inner crater, which is offset from the outer crater.

Therefore, we interpret the inner crater as the source of the rays and secondary craters of interest, and younger than the most recent episode of latitude-dependent mantling deposition at this low latitude. Our morphological observations suggest that the outer crater predates the end of

an obliquity-controlled period of latitude-dependent mantle deposition, while the inner crater appears to postdate the most recent period of mantle deposition (Head et al., 2003). To test this proposition quantitatively, we performed crater counts on smooth near-rim units of the inner crater. These units north and south of the inner crater both yield crater retention ages of ca. 1.25 Ma (Fig. 3B), based upon isochrons of Hartmann (2005). Some uncertainty exists in the production rate of craters at this size range; however, recent recognition of small craters (Malin et al., 2006) has provided observational evidence that these inferred recent cratering rates on Mars are unlikely to be off by more than a factor of a few (Hartmann, 2007; Kreslavsky, 2007). Thus, including the inferred uncertainty in production rates, the age range for the chronostratigraphic marker is between 0.6 Ma and 2.4 Ma.

Discussion and Conclusions

This study has identified a gully system depositional fan in eastern Promethei Terra containing secondary craters that are chronostratigraphic markers in the deposition of the fan. Fan morphology indicates that multiple periods of activity were required for its construction and that the secondary craters were emplaced during an intermediate period in fan formation. The presence of multiple superposing crater-free lobes requires several episodes of gully activity postdating emplacement of the secondary craters. Therefore,

the emplacement of the secondaries provides a firm maximum age on the most recent activity of this gully system. Approximately 100 km to the southwest, a 7-km-diameter rayed crater was identified that is interpreted to be the source of the secondaries.

Impact crater size-frequency distributions (Fig. 3B) place this crater's formation in the waning stages of the most recent period of latitude-dependent mantle accumulation and modification (Fig. 3C) (Head et al., 2003). The higher-amplitude obliquity variations during this period favor both the deposition and top-down melting of ice-rich deposits amenable to gully formation (Costard et al., 2002; Head et al., 2003). These stratigraphic relationships imply that at least some gullies on Mars have been active in very recent periods of the Late Amazonian during recent ice ages (e.g., Head et al., 2003; Schorghofer, 2007).

The multiple episodes of gully-related depositional fan activity mapped in this study imply that these gullies are not catastrophic landforms that formed in single events (i.e., as one-time debris flows or outbursts of groundwater). The distinctive alluvial fan-style morphology, fluvial channel sedimentary structures, and alcove incision make dry mass-wasting processes implausible for the formation of the gully system. The multiple episodes of activity required by the fan stratigraphy documented here cast doubt on deep groundwater discharge scenarios that are less likely to generate episodic releases. Rather, small amounts of surficial meltwater derived from snow and

ice accumulation are suggested by the insolation geometries of gully systems and can account most plausibly for multiple periods of recent activity required by these observations. Modeling by Williams et al. (2008) shows that martian snowpacks can reach melting temperatures under a variety of conditions and produce small amounts of meltwater. Freshly uncovered snowpacks under current conditions, snowpacks at higher obliquities, and windblown snow proposed by Head et al. (2008) to seasonally concentrate in gully channels, can all lead to small amounts of melt water (e.g., Christensen, 2003). These multiple avenues of surficial meltwater generation and our stratigraphic observations place recent gully activity during the most recent deposition and modification of latitude-dependent ice-rich mantling deposits.

Acknowledgements

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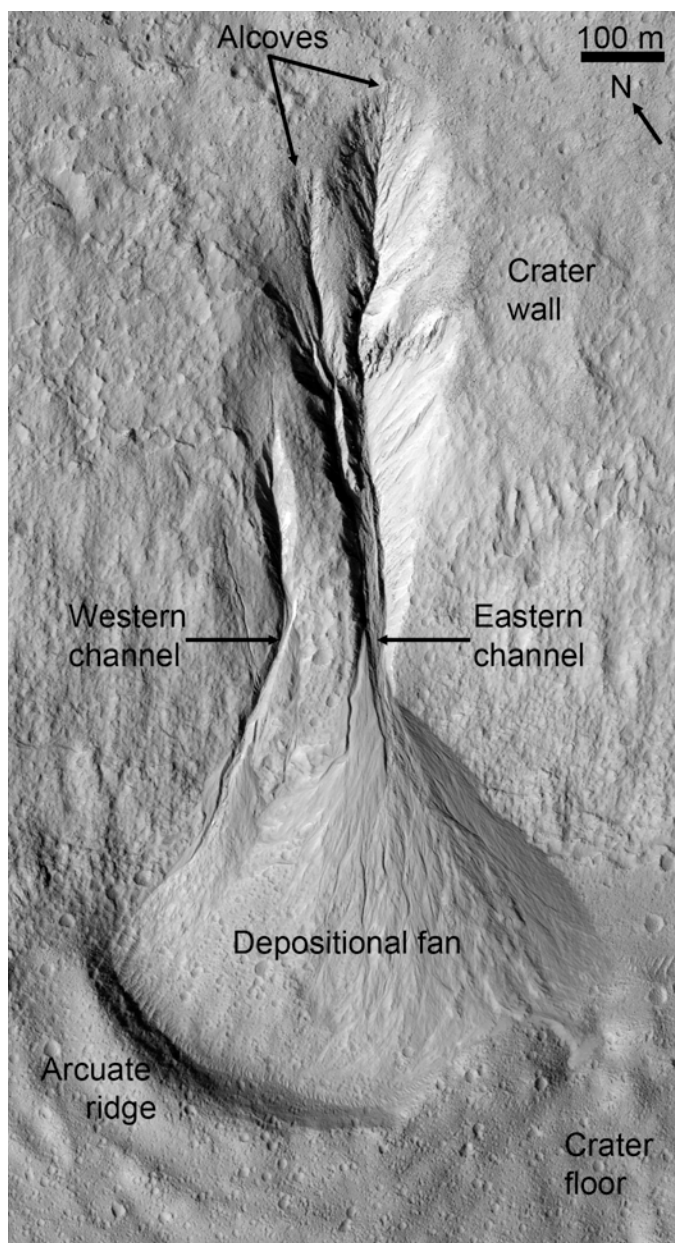


Figure 1. Eastern Promethei Terra crater wall gully system.

Tiered alcoves contribute to eastern and western channels feeding the depositional fan complex. Distinct lobes provide evidence for multiple episodes of gully activity (HiRISE: PSP_002293_1450).

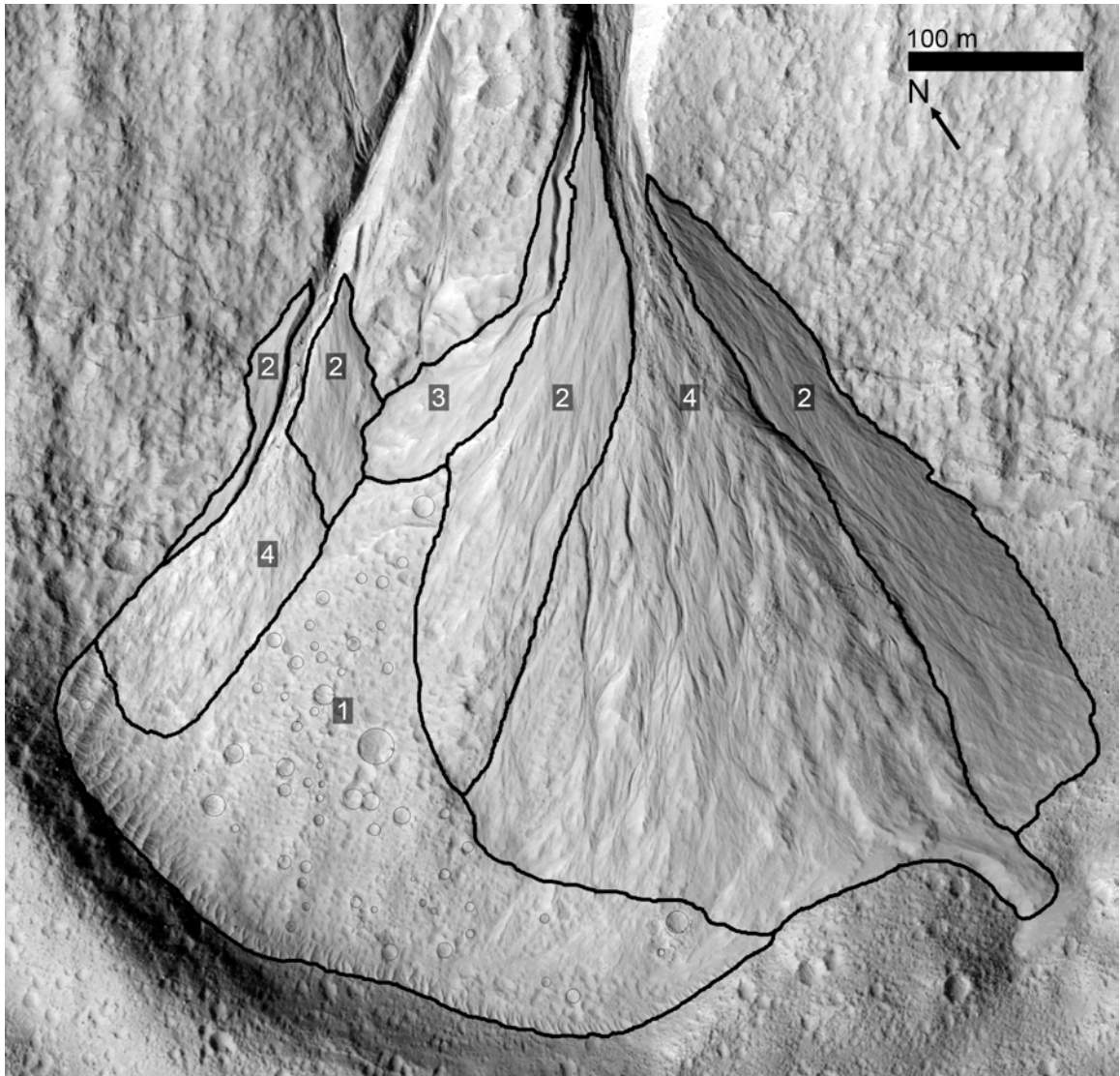


Figure 2. Sketch map of gully system depositional fan. The depositional fan is composed of six visible lobes. Lobe 1 is oldest visible lobe and retains dense population of secondary craters. Superposing uncratered lobes (2–4) postdate emplacement of secondary craters and require episodes of more recent gully activity (HiRISE: PSP_002293_1450).

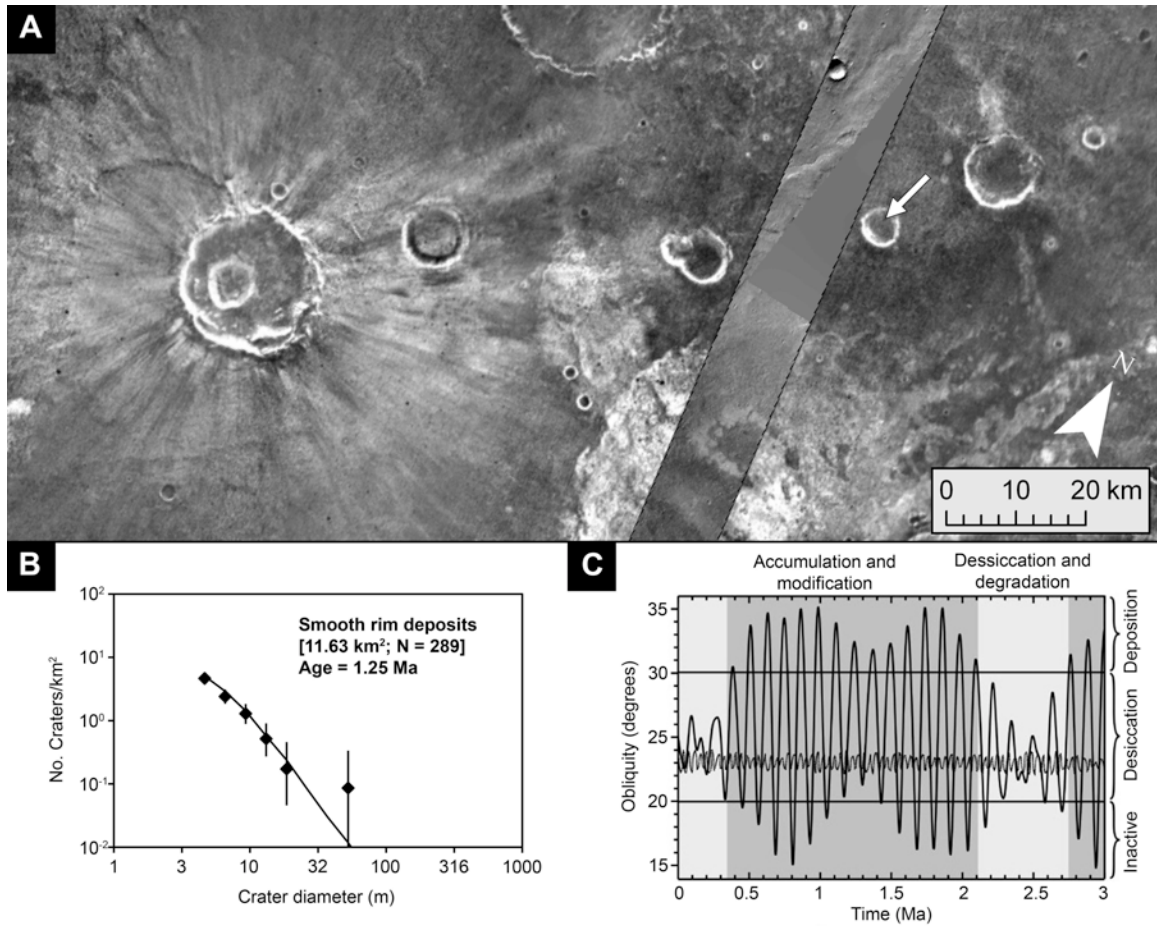


Figure 3. Eastern Promethei Terra (35°25'S, 130°25'E). A: THEMIS (Thermal Emission Imaging System) nighttime thermal infrared images show a pattern of fresh rays emanating from the ~7-km-diameter inner crater. The crater containing fan deposits (Figs. 1 and 2) is highlighted with white arrow. B: Crater counts displayed on incremental size-frequency plot of smooth near-rim deposits of the inner crater yield a crater retention age of ca. 1.25 Ma, placing formation of this crater in a period of obliquity-controlled mantle accumulation and modification. C: Mars obliquity variations (Laskar et al., 2002) over the past 3 Ma, with periods of mantle accumulation and modification (dark gray) and desiccation and degradation (light gray)

indicated. Low-amplitude line between 22° and 24° is the obliquity range of Earth (after Head et al., 2003).

CHAPTER 3

KEYS TO GULLY FORMATION PROCESSES ON MARS: RELATION TO CLIMATE CYCLES AND SOURCES OF MELTWATER.

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Abstract

Advances in dating gullies on Mars using superposition relationships and a stratigraphic marker horizon link gully chronostratigraphy to recent climate cycles. New observations of gully morphology show the close association of gully source regions, channels, and fan deposits with well-documented ice-rich latitude-dependent mantle deposits, the deposition of which is interpreted to be coincident with recent ice ages. On the basis of these close correlations, we interpret the formative processes for mid-latitude gullies to involve melting of these ice-rich mantling deposits and the generation of an aqueous phase leading to fluvial activity. Continued monitoring of gullies by spacecraft in the current “interglacial” climate period (~0.4 Ma to the present) will permit assessment of changing rates and styles of gully activity in the now largely depleted source areas.

Introduction

Originally discovered in images taken by the Mars Orbiter Camera, gullies on Mars were initially hypothesized to be the result of groundwater outbursts (Malin and Edgett, 2000). Additional hypotheses proposed alternative sources of water to carve gullies (e.g., ground ice; Costard et al., 2002; “pasted-on” terrain; Christensen, 2003), as well as entirely dry mechanisms for their formation (e.g., Shinbrot et al., 2004). The global distribution of gullies (Dickson et al., 2007; Dickson and Head, 2009), specific geologic studies (Christensen, 2003; Schon et al., 2009a; Levy et al., 2010; Morgan et al., 2010), terrestrial analog studies (Arfstrom and Hartmann, 2005; Head et al., 2007) and modeling efforts (Costard et al., 2002; Williams et al., 2009) favor variations of a meltwater scenario (e.g., Head et al., 2008) for the formation of gullies. New very high-resolution (sub-meter) image data from HiRISE provide striking details of gully characteristics and led McEwen et al. (2007) to report “evidence of fluvial modification of geologically recent mid-latitude gullies.” With these studies supporting a prominent role for water in forming martian gullies, ongoing research efforts are focused on (1) constraining the timing of gully formation (Reiss et al., 2004, 2010; Malin et al., 2006; Schon et al., 2009a; Dundas et al., 2010), (2) investigating specific formation processes (e.g., Pelletier et al., 2008; Kolb et al., 2010a; Levy et al., 2010), (3) exploring linkages between gully processes and inferred climate cycles (Mustard et al., 2001; Head et al., 2003; Laskar et al., 2004;

Schorghofer, 2007), and (4) determining candidate sources for meltwater (e.g., groundwater, ground ice, perennial ice, snow, and older glacial deposits) that may have been involved in fluvial activity (Head et al., 2003, 2007). Here we outline how new and recent observations address these questions by considering gullies in chronostratigraphic context and by demonstrating a close association between gully formation processes and meltwater derived from recent (<5 Ma) ice age deposits.

Dating Gully Formation

At the time of their discovery gullies were recognized as “geologically young” owing to a conspicuous absence of superposed impact craters or degraded morphologies (Malin and Edgett, 2000). Likewise, Malin and Edgett (2000) also describe in their report “other properties that similarly suggest relative youth, including superposition of aprons on eolian bedforms in Nirgal Vallis, superposition on polygonally patterned ground and the absence of rejuvenated polygons.”

It was difficult to advance beyond this basis of gullies as “geologically young” because conventional techniques of calculating crater retention ages for planetary surfaces are not robust on the limited areas and high slopes of gully environments. Reiss et al. (2004) were the first to date gully development using a superposition relationship. Depositional fans from gullies in the pole-facing wall superpose the transverse dune population in

Nirgal Vallis, which they report to have a crater retention age of 300,000 years to 1.4 Myr (Fig. 1). Reiss et al. (2004) concluded, “The last phase of more than 30°-obliquity at around 400,000 years [Laskar et al., 2004] correlates with the best fit model ages around 300,000 years for dune activity...Therefore gullies must have been formed after the last active phase of the dunes and are younger than 3 Myr, possibly less than 300,000 years.”

In eastern Promethei Terra, Schon et al. (2009a) identified a well-developed gully system in a crater wall that they were able to date using a novel technique utilizing the emplacement of a secondary crater population as a chronostratigraphic marker. By identifying and dating the rayed source crater of these secondary craters (Fig. 1) Schon et al. (2009a) showed, “Multiple lobes that post-date the secondary crater population make the emplacement date [0.6–2.4 Ma; best fit: 1.25 Ma] of the secondary craters a robust maximum age for the youngest lobes of this fan, and therefore the most recent activity of the gully system.” Schon et al. (2009a) concluded, “The presence of multiple superposing crater-free lobes [of the depositional fan] requires several episodes of gully activity postdating emplacement of the secondary craters. Therefore, the emplacement of the secondaries provides a firm maximum age on the most recent activity of this gully system” (Figs. 1 and 2).

The concurring observations of both Reiss et al. (2004) and Schon et al. (2009a) are interpreted similarly with respect to recent obliquity-driven

(Laskar et al., 2004) climate cycles (Fig. 1) and can be linked to the climate conditions that are thought to have prevailed at that time. Specifically, on the basis of a wide variety of evidence, the period of enhanced obliquity from ~0.4 to 2.1 Ma (Fig. 1) has been interpreted to represent a “glacial” period or “ice age” during which ice-rich layers were deposited from about 30° north and south latitudes to the poles in the form of a many meters-thick mantle (e.g., Head et al., 2003; discussed further below). In the current period of lower amplitude obliquity variations (Fig. 1), the latitude-dependent mantle is undergoing degradation in lower latitude portions (~30–50°N and S latitude). These relationships suggest that gully chronostratigraphy (Fig. 1) can provide important links to climate history and the processes that form and modify gullies. Here we use new observations of alcove-less gullies to suggest a candidate source of water for gully formation using chronostratigraphic constraints.

Gullies in a Stratigraphic Context

The locale in Fig. 2 (~35°S, 131°E) provides an informative setting for an analysis of the stratigraphic position of gullies in relation to the deposition and erosion of these widespread ice-rich mantling deposits. At this location, secondary impact craters from the formation of Gasa crater established a stratigraphic marker horizon, which can be used to constrain gully timing and coincident degradation of mantling materials. Stratigraphic markers,

such as these secondary craters, are important tools in chronostratigraphy because they represent a layer or event that was simultaneously emplaced over a wide area in different depositional environments; such markers on Earth include large volcanic eruptions that produce ash/ bentonite marker beds and geochemical-stratigraphic markers formed by the widespread deposits of impact cratering events. Gasa secondary craters (Fig. 2) are pervasive on the crater floor and rim, but are much less well-developed, degraded, or absent on the crater wall and much of the gully fan deposits.

What is the broader context of this stratigraphic marker on Mars? Mid- to high- latitude geomorphology of latest Amazonian age on Mars is characterized by ice-related processes and landforms including a pervasive ice-rich mantling unit first identified in global maps of surface roughness (Kreslavsky and Head, 2000) that show topographic smoothing at high latitudes; evidence for this mantle is also seen in visual imaging (Mustard et al., 2001; Kreslavsky and Head, 2002; Milliken et al., 2003). Head et al. (2003) synthesized these observations into a theory of recent obliquity-driven “ice ages” on Mars that is supported by global circulation model studies (Mischna et al., 2003; Levrard et al., 2004) and models of ice stability (Schorghofer, 2007). Additional evidence supporting extensive atmospheric deposition of ice is provided by Gamma Ray Spectrometer (GRS) data of ice abundances that far exceed reasonable pore space volumes (Boynton et al., 2002), observations by Phoenix of massive ice just below the surface (Smith et

al., 2009), observations of sublimation-type contraction crack polygons at the Phoenix landing site (Levy et al., 2008), observations of layering within the mantling unit (Schon et al., 2009b), and contemporary observations of new mid-latitude impact craters which expose a nearly pure ice substrate that is observed to sublime upon exposure (Byrne et al., 2009; Dundas and Byrne, 2010). The formation of the stratigraphic marker is clearly coincident with the period of time during which this ice-rich mantling deposit was emplaced at these latitudes (Fig. 1).

Degradation of the latitude-dependent mantle

Analysis of the crater floor, wall, and rim (Fig. 2A) show the presence of the regional latitude-dependent ice-rich mantle on the floor and crater rim; these areas are also characterized by pervasive superposed secondary craters from Gasa crater (e.g., Schon et al., 2009a) (Fig. 2A, top and bottom), a relationship that dates much of this mantle emplacement to pre-Gasa history (Fig. 1). Also observed on the crater wall are several scarps that face upslope, and have very irregular, sinuous to digitate borders. The most prominent example occurs near the base of the crater wall and consists of a generally continuous scarp facing upslope with a sinuous outline (right-hand side of Fig. 2A). Examination of the broad crater wall reveals an additional but less distinctive scarp in the middle part of the crater wall (Fig. 2A). The boundary of this scarp is much more digitate than the boundary of the scarp at the base

of the wall; indeed, examination at higher resolution shows that this scarp is composed of two to three smaller digitate scarps (Fig. 2D, top). In the upper part of the crater wall, two or three additional scarp-like trends are seen, some relatively continuous for hundreds of meters, and others less continuous (see Fig. 2A, upper left in particular). We interpret these scarps to be the eroded remnants of the multiple layers shown to comprise the latitude-dependent mantle and documented in many places elsewhere in areas where the mantle is undergoing erosion (e.g., Schon et al., 2009b). In this scenario, the wall surface represents an area of mantle undergoing degradation, and the scarps represent the exposed remnants of mantle layers. The stratigraphic relationships indicate that the layers higher up on the wall represent progressively older depositional layers.

Detailed analysis of the state of preservation of the pervasive secondary craters from Gasa crater in this area provides further insight. Secondary craters on the floor are well preserved (Fig. 2A, bottom; D, bottom), but secondaries lying above the scarp at the base of the crater wall appear more irregular and incomplete (Fig. 2C, middle; D, middle), consistent with their modification by erosion and degradation of the mantle layers. Areas higher on the crater wall display even more degraded secondary craters (Fig. 2C, top; D, top) or little evidence of secondaries at all (Fig. 2A, middle and top), while other orientations of the crater wall preserve a smooth mantle texture and many secondaries (Fig. 2A, far right). Analysis of the

stratigraphic relationships of the gully fan deposits and the mantle layers modified by secondaries from Gasa (Fig. 2A; B, bottom; C, bottom; D, bottom) shows that portions of the fans are clearly superposed on, and thus post-date, the emplacement of secondaries.

Together, these observations are consistent with (1) the region being mantled by an ice-rich latitude-dependent deposit, (2) pole-facing crater-wall slopes having the ice-rich mantle preferentially degraded, (3) the scarps representing the margins of layers composing this progressively degraded mantle, (4) the mantle undergoing significant degradation following the Gasa crater-forming event that emplaced the secondary craters in this region, and (5) gully activity (portions of gully fans) continuing to occur following the Gasa cratering event (Fig. 1).

The primary crater that is the source of the secondary craters superposed on the mantle (and thus representing a marker event in the region) is the Gasa impact crater, located ~100 km to the southwest. The crater size frequency distribution on Gasa ejecta indicates that this event occurred ~1.25 Ma with uncertainty in the production function for small craters yielding a range of 0.6–2.4 Ma (Schon et al., 2009a). Unambiguously superposing the stratigraphic marker are portions of the gully fans (Fig. 2A).

Alcove-less gullies suggest a candidate source of meltwater

Could the ice-rich latitude-dependent mantle be a source of water for formation of these gullies? Examination of HiRISE data helps to address this question. The most prominent gully system occurs in the central part of the crater wall (Fig. 2A and C). Smaller surficial gullies also occur, however, in the degraded mantle on the crater wall (Fig. 2B and D). These features lack alcoves, have shallowly incised channels, and have depositional fans that extend onto the cratered floor surface (Fig. 2A). Furthermore, shallow channels emanating from the mantle feed into the larger gully system (Fig. 2C, arrows 1–3) and occur independently nearby the larger gully (Fig. 2C, arrow 4). In Fig. 2B, the uppermost narrow channels (arrows 1 and 2) are choked with sediment (arrow 3) above a more deeply incised portion of the channel that is also choked with sediment (arrow 4); below this, a channel segment (arrow 5) and subsequent deposition (arrow 6) suggest multiple sediment transport episodes in the development of this gully. In Fig. 2D, the upper reaches (arrows 1 and 2) of the channel are shallower and appear restricted by the boundary of a mantle layer (Fig. 2A). The occurrence of these small-scale surficial gullies and channels in the degraded mantle, without alcoves that could serve as accumulation zones for snow, provides independent evidence that gullies can form through degradation and melting of the ice-rich mantling deposits. This is consistent with interpretations (e.g., Head et al., 2003) of the correlation between gully activity and obliquity-driven climate cycles (Fig. 1).

In summary, the intimate association of the degrading portions of the ice-rich latitude-dependent mantle with the sources of gullies on the crater wall, and the similarities of gully channels and fans to features formed by liquid water-related fluvial activity (e.g., Head et al., 2007, 2008; McEwen et al., 2007), strongly favor a fluvial origin for many of the major gully features sourced from melting ice-rich mantle material. Modeling the stability of snow and buried snow and ice, Williams et al. (2009) showed that melting in these latitude bands could take place during peak insolation geometries for a part of the year. Furthermore, the chronostratigraphic marker links the earliest time of this gully activity to the latter part of the recent glacial period (Fig. 1). On the basis of these analyses, we hypothesize that the steep slopes of the crater wall preferentially orient the ice-rich mantle deposits so that peak insolation trends during obliquity excursions (Fig. 1) (similar to the peak insolation geometries described by Costard et al. (2002)) cause melting of ice (e.g., in a manner similar to that envisioned by Williams et al. (2009)). The liquid water liberated by melting of ice during periods of peak insolation resulted in multiple phases of fluvial activity that formed the gullies. Continued ablation (melting and sublimation) resulted in the degradation and removal of much of the ice-rich mantle deposit.

Activity during the current “interglacial” period

Gasa crater, source of the pervasive secondaries, also hosts multiple gullies, which must post-date the impact event. On the basis of the chronostratigraphy, these gullies can provide insight into the nature of gully activity in the waning stages of the “glacial” period and into the current “interglacial.” Kolb et al. (2010b) described the Gasa crater gullies as the best preserved (freshest morphological appearance), relative to gullies of their other study areas; they investigated gully development by analyzing the gradient where deposition begins (apex slope of the fan) using digital elevation models derived from HiRISE stereo pairs. Analysis of Gasa crater gullies by Kolb et al. (2010b) shows that ten gullies have apex slopes characteristic of “wet or fluidized emplacement” (16.3–20.4°) and eleven gullies have apex slopes consistent with dry granular flows (20.7–26.4°). These data are consistent with the Gasa impact occurring prior to the likely waning of meltwater generation for gully formation, which we suggest is contemporaneous with the damping of obliquity-climate cycles, ~400 ka (Fig. 1). Kolb et al. (2010b) also found that the morphologically freshest appearing gullies were more likely to have apex slopes consistent with the movement of dry material, while more degraded gullies were more likely to have slopes consistent with wet flows. “Our results suggest that gully formation required a time-limited fluidization mechanism, possibly liquid water, that major gully formation is not occurring today, and that activity in gullies today is likely dry mass wasting perhaps aided by CO₂ frost,” concluded Kolb et al. (2010b).

Dundas et al. (2010) detected recent rockfalls and other minor geomorphic changes in two Gasa gullies and suggested a link between the annual CO₂ frost cycle and these events, but concluded, “None of these observations contradict the hypothesis that gullies are initiated by H₂O snowmelt or that this process drives a significant fraction of gully erosion.” In summary, these observations suggest that the role of liquid water in gully activity may have become less important in the transition to the current interglacial period, as environmental conditions became less extreme and ice-rich mantling deposits became depleted.

Conclusions and Implications

New chronostratigraphic data and observations have improved our understanding of the processes and climate context for the formation and evolution of martian gullies. While mid-latitude gullies on Mars are geologically young features dating to the latest Amazonian, their principal formation is likely to have preceded the current (0–400 Kyr) epoch of lower and more stable obliquity (Fig. 1). In our interpretation, melting of ice-rich deposits related to degradation of a latitude-dependent mantle was responsible for a major portion of the fluvial activity forming gully channels and depositional fans (Fig. 2). Erosive gully-forming flows are likely to have been dominated by fluvial sediment transport (e.g., McEwen et al., 2007, 2010; Head et al., 2007; Schon et al., 2009a), with a few cases of water-

lubricated debris flows (e.g., Levy et al., 2010; Lanza et al., 2010). In contrast, present-day gully activity may favor dry mass movements, which appear to exhibit a seasonal modulation related to the CO₂ frost cycle (Dundas et al., 2010; Diniega et al., 2010). Seasonal monitoring and ongoing change detection campaigns (e.g., McEwen et al., 2010) will be essential to investigate active processes and to characterize geomorphic changes on steep slopes and in gullies.

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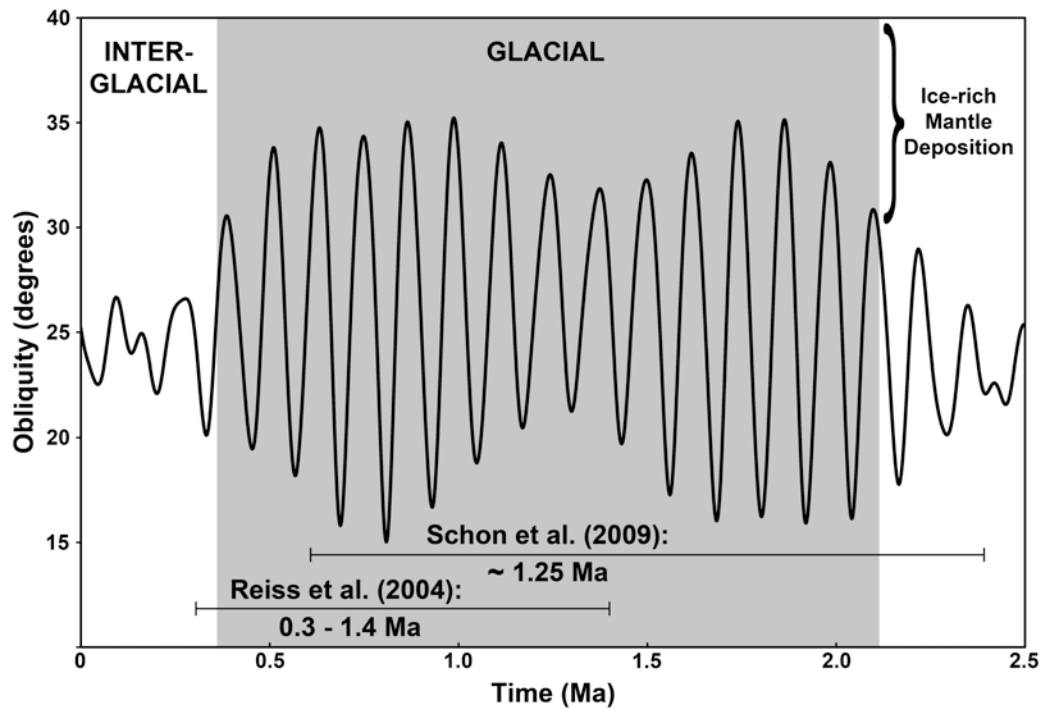


Figure 1: Obliquity variations of Laskar et al. (2004) and suggested climate epochs of Head et al. (2003). Higher obliquity and higher amplitude variations in the recent past (0.4– 2.1 Ma) define episodes of mid-latitude ice stability (e.g., Mellon and Jakosky, 1995) during which the deposition of multiple ice-rich mantling layers has been interpreted from geological evidence. Meltwater derived from degradation of this mantling unit is interpreted to be one source of meltwater responsible for gully formation, along with later snow in gullies. This chronology is consistent with gully dating via superposition and chronostratigraphic markers that provide maximum gully ages (Reiss et al., 2004; Schon et al., 2009a,b).

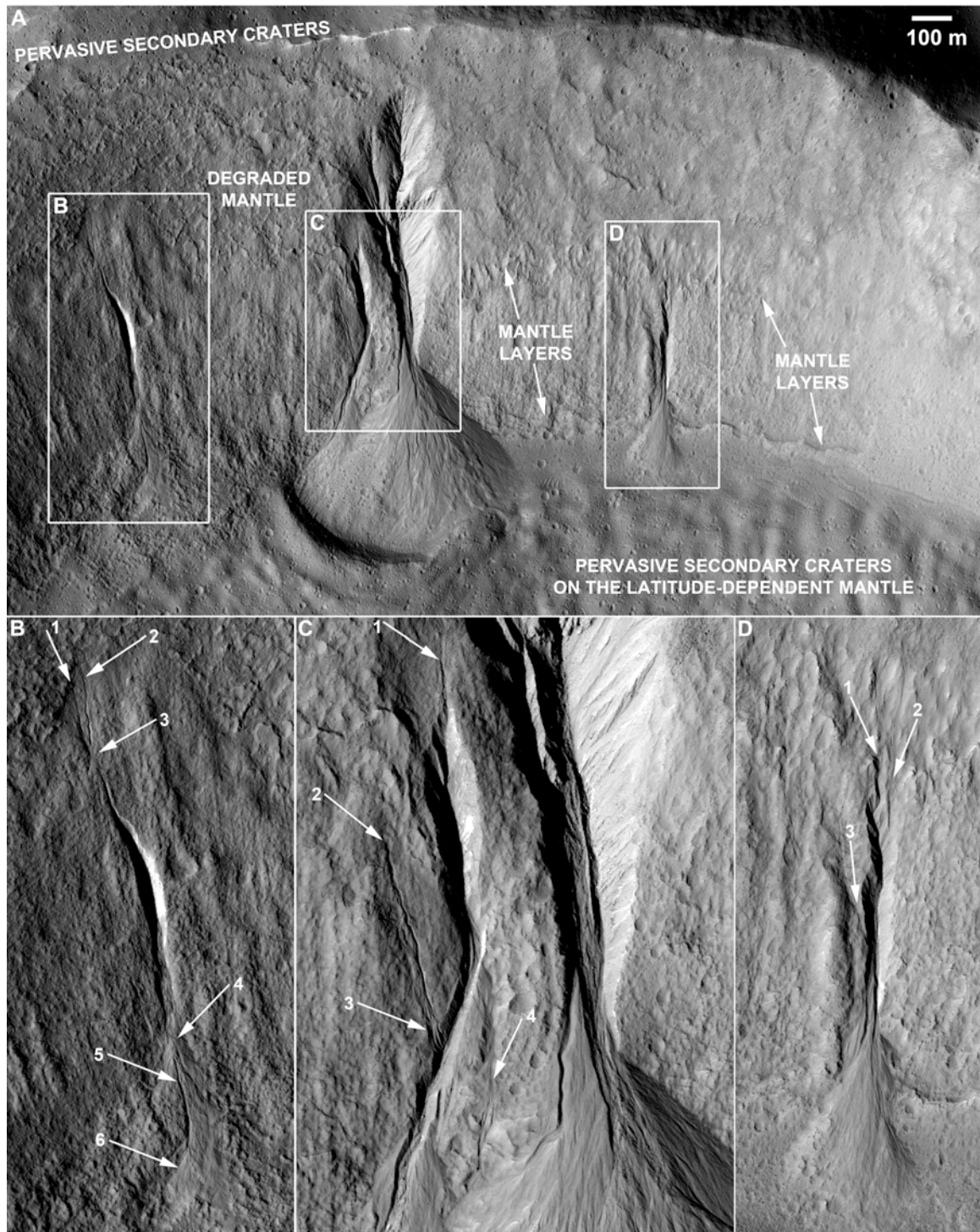


Figure 2: Gully depositional fans in a 5-km crater (A) post-date a dense population of secondary craters (eastern Promethei Terra, $\sim 35^{\circ}\text{S}$, 131°E). The pole-facing wall (shown) is composed of degraded layers of ice-rich latitude-

dependent mantling deposits. On the crater wall, degradation of the mantling unit has led to a partial to nearly complete removal of the secondary crater population and exposed prominent layers of the mantle with scarps at the margins. Small channels and gullies without alcoves emerge from the degraded mantle and are interpreted to have deposited fans during degradation and melting of the ice-rich mantling material. Fine-scale surficial gully features are highlighted with arrows in insets (B–D) as described in the text. These observations implicate meltwater from the degradation of latitude-dependent mantling in the process of gully formation. HiRISE PSP_002293_1450.

CHAPTER 4

GASA IMPACT CRATER, MARS: CHRONOLOGY OF GULLY DEVELOPMENT AND DERIVATION OF MELTWATER FROM LATITUDE DEPENDENT MANTLE AND EXCAVATED DEBRIS-COVERED GLACIER DEPOSITS.

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Abstract

The mode of formation of gullies on Mars, very young erosional-depositional landforms consisting of an alcove, channel and fan, is one of the most enigmatic problems in martian geomorphology. Major questions center on their ages, geographic and stratigraphic associations, relation to recent ice ages, and, if formed by flowing water, the sources of the water to cause the observed erosion/deposition. Gasa (35.72°S, 129.45°E), a very fresh 7-km diameter impact crater, and its environment offer a unique opportunity to explore these questions. We show that Gasa crater formed during the most recent glacial epoch (2.1-0.4 Ma), producing secondary crater clusters on top

of the latitude-dependent mantle (LDM), interpreted to be a layered ice-dust-rich deposit emplaced during this glacial epoch. High-resolution images of a pre-Gasa impact crater ~100 km northeast of Gasa reveal that portions of the secondary-crater-covered LDM have been removed from pole-facing slopes in crater interiors near Gasa; gullies are preferentially located in these areas and channels feeding alcoves and fans can be seen to emerge from the eroding LDM layers to produce multiple generations of channel incision and fan lobes. We interpret these data to mean that these gullies formed extremely recently in the post-Gasa-impact time-period by melting of the ice-rich LDM. Stratigraphic and topographic relationships are interpreted to mean that under favorable illumination geometry (steep pole-facing slopes) and insolation conditions, melting of the debris-covered ice-rich mantle took place in multiple stages, most likely related to variations in spin-axis/orbital conditions. Closer to Gasa, in the interior of the ~18 km diameter LDM-covered host crater in which Gasa formed, the pole-facing slopes display two generations of gullies. Early, somewhat degraded, gullies have been modified by proximity to Gasa ejecta emplacement, and later, fresh appearing gullies are clearly superposed, cross-cut the earlier phase, and show multiple channels and fans, interpreted to be derived from continued melting of the LDM on steep pole-facing slopes. Thus, we conclude that melting of the ice-rich LDM is a major source of gully activity both pre-Gasa crater and post-Gasa crater formation. The lack of obscuration of Gasa secondary clusters

formed on top of the LDM is interpreted to mean that the Gasa impact occurred following emplacement of the last significant LDM layers at these low latitudes, and thus near the end of the ice ages. This interpretation is corroborated by the lack of LDM within Gasa. However, Gasa crater contains a robustly developed set of gullies on its steep, pole-facing slopes, unlike other very young post-LDM craters in the region. How can the gullies inside Gasa form in the absence of an ice-rich LDM that is interpreted to be the source of water for the other adjacent and partly contemporaneous gullies?

Analysis of the interior (floor and walls) of the host crater show that prior to the Gasa impact, the pole-facing walls and floor were occupied by remnant debris-covered glaciers formed earlier in the Amazonian, and relatively common in crater interiors in this latitude band. We show that the Gasa impact cratering event penetrated into the southern portion of this debris-covered glacier, emplaced ejecta on top of the debris layer covering the ice, and caused extensive melting of the buried ice, and flow of water and debris slurries on the host crater floor. Inside Gasa, the impact crater rim crest and wall intersected the debris-covered glacier deposits around the northern, pole-facing part of the Gasa interior. We interpret this exposure of ice-rich debris-covered glacial material in the crater wall to be the source of meltwater that formed the very well-developed gullies along the northern, pole-facing slopes of Gasa crater.

Introduction

Martian gullies, defined by the presence of an alcove, channel, and fan, were discovered in the first meters-scale images of Mars (Malin and Edgett, 2000). While provocative, significant uncertainty surrounded their formation, history, and potential link to climate cycles. Some initial hypotheses for gully formation depended only on dry mass wasting and granular flows (Treiman, 2004; Shinbrot et al., 2004) which would be largely climate agnostic, but most hypotheses imply that some form of aqueous fluidization mechanism is required. For example, wet gully formation scenarios that have been proposed include groundwater outbursts from shallow aquifers (Malin and Edgett, 2000; Heldmann and Mellon, 2004), deep groundwater outbursts (Mellon and Phillips, 2001), melting of shallow ground ice (Costard et al., 2002; Hartmann et al., 2003), melting of older snowpacks (Christensen, 2003), and similar melting of surficial snow and ice accumulations (Head et al., 2007; Head et al., 2008; Dickson and Head, 2009; Williams et al., 2009).

While recent work strongly shows that mid-latitude gully formation required an aqueous phase (e.g., McEwen et al., 2007b), uncertainty is only now being resolved regarding the chronology of gully development, the timescale required for their formation, and variations in the style of sediment transport with time. These characteristics are crucial to evaluation of latest Amazonian climate and of gullies as a geomorphic process on Mars. Additionally, improved understanding of gully development will have

important implications regarding the most recent occurrences of liquid water, which is of interest to the astrobiology research community (e.g., MEPAG, 2006; Space Studies Board, 2007).

Reiss et al. (2004) were the first to date gully deposition via superposition relationships in Nirgal Vallis. Schon et al. (2009a) showed concurring chronostratigraphic data derived from a secondary crater population marker horizon indicating significant gully deposition postdating ~ 1.25 Ma. These marker horizons suggest a direct link between gully formation and the degradation of surface ice and snow deposits related to latest Amazonian climate cycles (Schon and Head, 2011). Here we present a study of Gasa crater for which excellent absolute and relative age dating of the crater allows us to consider the end-to-end process of gully development and evolution with temporal constraints. Impact into an ice-rich substrate contributed significantly to the development of gullies in Gasa crater. Gasa crater occurs within a larger 18-km diameter crater that hosts gullies and other geomorphic evidence of significant glacial ice accumulation. Nearby craters contain evidence of similar ice accumulations. The development of large gullies within youthful Gasa crater underscores the remarkable efficiency of gully formation as a geomorphic process.

Geologic Setting

Gasa crater (35.72°S, 129.45°E) (Fig. 1) is a very fresh ~7-km diameter impact crater which occurs within the simple to complex crater transition on Mars (Pike, 1980; Garvin et al., 2003) with a sharp rim-crest, a well-defined flat floor, and no evidence of substantive infilling (depth-diameter ratio of 0.12). Gasa crater is located within an older ~18-km diameter crater in eastern Promethei Terra on Noachian cratered terrain (Fig. 1). The outer (unnamed) host crater has muted topography, a low depth:diameter ratio (0.07), evidence of latitude-dependent mantling (e.g., Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003), gullies, and polygonally patterned ground (e.g., Levy et al., 2009a), which are all evidence of extensive Amazonian modification (e.g., Kreslavsky and Head, 2006).

Crater Rays: Radiating from Gasa crater are extensive ray patterns (Fig. 1A) visible in nighttime THEMIS infrared data (Schon et al., 2009). The population of rayed craters on Mars, such as Gasa crater, is limited (Tornabene et al., 2006). Generally, the distinctiveness of crater rays arises from both compositional and maturity differences (e.g., Hawke et al., 2004). The thermophysical distinctiveness of martian rays (McEwen et al., 2005; Tornabene et al., 2006; Preblich et al., 2007) is attributable to thermal inertia (TI) differences with surrounding terrain (e.g., low TI rays). Hence, rays are most apparent in nighttime THEMIS IR data (Christensen et al., 2004). The distribution of identified rayed craters (Tornabene et al., 2006) suggests that the occurrence or persistence of rays is dependent on substrate. Intermediate

to high background thermal inertia and intermediate albedo appear to be important criteria regarding the distinguishability of rays (see global maps of Mellon et al., 2000 and Putzig et al., 2005). This is consistent with most of the Tornabene et al. (2006) detections occurring in volcanic terranes. Rays on Mars can be homogenized with the surrounding environment or obscured from view by multiple processes known to be active on the martian surface, including: glacial modification, dust deposition, eolian reworking, and volcanic flows. For example, rays from Zunil are conspicuously absent from the Medusae Fossae Formation (Preblich et al., 2007), which Kerber and Head (2010) have shown to have a complex history of erosion and reworking. Gasa crater is located farther poleward (35.7°S) than any rayed crater identified by Tornabene et al. (2006). The apparent dearth of rayed craters in the mid and high latitudes is consistent with recent resurfacing events in these regions due to emplacement of latitude-dependent mantling deposits (Head et al., 2003) discussed further below (Schon et al., 2011).

Crater Age: The Gasa crater-forming impact event created a smooth near-rim ejecta deposit. This deposit is ideal for crater-retention age dating because of its smooth surface and gentle topography. Using HiRISE data (McEwen et al., 2007a), 289 craters were identified on 11.63-km² of the smooth ejecta deposit with the largest crater 61 m in diameter (Schon et al., 2009a). Employing isochrons of Hartmann (2005), this crater size-frequency distribution implies a best-fit age of 1.25 Ma for the Gasa crater impact event

(Fig. 2). While there is some uncertainty in the production rate of small craters, recent direct observations of small crater formation (Malin et al., 2006) suggest that inferred recent cratering rates are unlikely to be in error by more than a factor of a few (Hartmann, 2007; Kreslavsky, 2007). Therefore, we have confidence in an age range of 0.6 to 2.4 Ma for the formation of Gasa crater (Schon et al., 2009a), which is entirely consistent with preservation of the rays (McEwen et al., 2005; Tornabene et al., 2006).

Ice-Rich Mantling Deposits

Mars' mid- to high-latitude (greater than $\sim 30^\circ$ North and South) latest Amazonian geomorphology is characterized by ice-related processes and landforms, including a pervasive ice-rich mantling unit observed in the larger crater encompassing Gasa crater as well as on the surrounding terrain. The mantling unit, responsible in part for early observations of terrain softening (e.g., Squyres and Carr, 1986), was first identified in global maps of surface roughness (Kreslavsky and Head, 2000) derived from MOLA altimetry, which reveal topographic smoothing at high latitudes. The morphology and degradation characteristics of this ice-rich unit have also been described from visual observations by Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003; Milliken et al., 2003; Milliken and Mustard, 2003; Kostama et al., 2006; and Morgenstern et al., 2007.

Ice Content: While the process of vapor diffusion governs the stability of ground ice deposits (e.g., Mellon and Jakosky, 1995), geological evidence suggests that the ice-rich mantling unit is the result of atmospheric deposition rather than vapor diffusion into regolith pore space. Evidence for extensive atmospheric deposition of ice is provided by Gamma Ray Spectrometer / Neutron Spectrometer data which imply ice abundances that far exceed reasonable pore space volumes for regolith (Boynton et al., 2002; Feldman et al., 2002; Mitrofanov et al., 2002; Prettyman et al., 2004). Also supporting the depositional hypothesis, Levy et al. (2008) have documented observations of sublimation-type contraction crack polygons at the Phoenix lander site (68.22°N, 234.25°E), which they interpret to require a nearly pure ice substrate. Similarly, the Phoenix lander observed massive ice beneath a thin regolith cover at this location (Smith et al., 2009). Finally, repeated observations have identified new mid-latitude impact craters that expose a nearly pure ice substrate, and this ice is observed to sublimate upon exposure (Byrne et al., 2009; Dundas and Byrne, 2010).

Depositional History: A theory of recent obliquity-driven “ice ages” was proposed by Head et al. (2003) that relates broad deposition of ice-rich mantling deposits to larger obliquity variations that occurred > 400 ka (Fig. 3). The validity of this scenario is supported by global circulation model studies (Mischna et al., 2003; Levrard et al., 2004), models of ice stability (Schorghofer, 2007), and evidence of layers within the mantling unit (Schon

et al., 2009b). This theory suggests that emplacement of the mantle was cyclical with depositional phases punctuated by periods of instability and degradation, as expected from the obliquity record (Fig. 3).

Relative Stratigraphic Position: While the precise chronological history of mantle emplacement is not yet resolved, initial efforts by Kostama et al. (2006) to directly date its uppermost surface in the northern hemisphere suggest that crater retention age trends correlate with latitudinal variations in morphology. The highest latitudes (70-80°N) exhibit the youngest ages (~0.1 Ma), while lower latitude mantle terrains appear older, ~2 Ma (Kostama et al., 2006). These trends are consistent with the preservation of rays from Gasa crater (Fig. 1A), the clear superposition of un-modified crater clusters and chains from Gasa crater on the latitude-dependent mantle surface (Fig. 4), and the lack of morphological evidence for mantle deposition within Gasa crater. In contrast, latitude-dependent mantling is present on the walls of the host crater. For example, gully alcove walls within the host crater exhibit polygonal texture (Fig. 5) indicative of the ice-rich substrate (e.g., Mangold, 2005; Levy et al., 2009a). The position and development of these polygons imply that some alcove development preceded the most recent ice-rich mantle deposition at these locations. Polygons are also observed on crater wall mantle surfaces near gully alcoves (Fig. 6). Multiple episodes of gully activity also post-date the mantle as evidenced by un-mantled gully deposits with multiple fans (Fig. 6). Surficial channels of limited incision that

are not associated with alcoves are also observed (Fig. 6) and provide evidence that melting of ice-rich LDM is sufficient for some gully activity (Schon and Head, 2011).

Ice-Rich Mantle and Gullies within the Host Crater: Gasa crater is located within the 18-km diameter host crater (Fig. 1). This crater contains latitude-dependent mantling deposits, hosts pole-facing gullies, and has a significantly lower depth-diameter ratio (0.07) than Gasa crater (0.12). The pole-facing wall of the host crater is modified by a series of gullies. The largest gully alcoves have widths of approximately 400 m to 500 m. Older fans are observed that are deformed by multiple closely spaced fractures parallel to the base of slope (e.g., Head et al., 2008), while stratigraphically younger fans are un-modified (Fig. 6). Younger fans are also located upslope and superpose previously eroded gully channels (Fig. 6). Alcoveless gullies, which are sourced directly from the mantle, (Fig. 6) similar to those described by Schon and Head (2011) and shown in Fig. 7 are also observed. The youngest activity is likely to be contemporaneous with similar activity that has been documented to post-date Gasa crater (Schon et al., 2009a; Schon and Head, 2011). Therefore, these observations support episodes of gully activity at this location that both predate (based on the presence of latitude-dependent mantling, which must predate Gasa, within a gully alcove, e.g. Fig. 5), as well as postdate Gasa crater.

Theories of Gully Formation

Melting of latitude-dependent mantling deposits has been proposed as a source of water for mid-latitude gullies (Head et al., 2003; Milliken et al., 2003; Bleamaster and Crown, 2005; Bridges and Lackner, 2006; Head et al., 2008; Dickson and Head, 2009; Schon and Head, 2011). From a process standpoint, this scenario is comparable to previously proposed sources of meltwater including ground ice (Costard et al., 2002) and ancient snowpacks (Christensen, 2003), as well as similar scenarios (e.g., Levy et al., 2009b; Lanza et al., 2010). It has also been proposed that gullies are late-stage features that develop during the wane and retreat of alpine-style glaciers (Arfstrom and Hartmann, 2005; Berman et al., 2005; Head et al., 2008). Evidence supporting this process model includes well-developed arcuate ridges interpreted as moraines below cirque-like alcoves that would have been ideal accumulation zones for glacial systems (Arfstrom and Hartmann, 2005; Berman et al., 2005; Head et al., 2008; Berman et al., 2009).

It is important to note that while these moraine features are prominently associated with some gullies, particularly in the east of Hellas region, they are not universally associated with mid-latitude gullies. In a recent study, Schon and Head (2011) documented small-scale surficial gullies whose incision is strictly limited to the mantling unit and which lack alcoves (Fig. 7). These gullies provide independent evidence that melting and degradation of ice-rich mantling deposits is an important source of

meltwater, sufficient for gully formation in some instances. At this locality, approximately 100 km northeast of Gasa crater, Schon et al. (2009a) mapped multiple gully deposits that superpose secondary craters from Gasa crater. These stratigraphic relations indicate that mantle degradation and gully activity at this site post-date the Gasa crater impact (Fig. 7).

Therefore, independent evidence exists for the development of gullies in association with two environments: 1) past glacial systems with commensurately large accumulations of ice during the late Amazonian (e.g., Berman et al., 2005; Head et al., 2008) and 2) from localized post-Gasa crater melting of latitude-dependent mantling deposits emplaced prior to Gasa (Schon and Head, 2011). Since these mantling deposits were emplaced pre-Gasa crater, how did the gullies within Gasa crater form? If water was involved, what was the source? The age, stratigraphic relationship with latitude-dependent mantling deposits, and geomorphic setting of Gasa crater enable us to directly address these questions.

Gasa Crater Interior

The rim crest of Gasa crater is crisply defined, but asymmetric in form. The northern portion is deeply crenulated by gully alcoves while the southern extent is composed of linear to curvilinear segments (Fig. 8). The alcoves are neither symmetric nor uniform. The largest alcoves are located in the middle of the northern rim (pole-facing). Alcoves are progressively smaller both

clockwise and counterclockwise from this orientation. Individual alcoves are complex, often with multiple contributing sub-alcoves that have associated channels. Sharp divides between alcoves expose fractured rocky material. An orientation asymmetry is also observed within alcoves. For example, the primary orientation of the alcoves shown in Fig. 9 is toward the south-southwest. In these alcoves, substantially more sub-alcoves and channels are found on the pole-facing walls. The pole-facing walls of these alcoves also have more bedrock exposures (Fig. 9). In contrast, the equator-facing, uncrenulated southern rim and wall of the crater is characterized by narrow poorly-developed debris chutes and bedrock exposures are comparatively sparse (Fig. 8). The debris chutes that have developed along the southern interior of Gasa crater are very immature (i.e., shallow and narrow) compared to those observed in Zunil, a larger ($D = 10.4$ km) young rayed crater (McEwen et al, 2005), but are similar to those observed in the other rayed craters identified by Tornabene et al. (2006) such as Zumba ($D = 3.3$ km) [PSP_003608_1510], Gratteri ($D = 6.9$ km) [PSP_010373_1620], and Tomini ($D = 7.4$ km) [PSP_001871_1965].

Gully alcoves and channels: Within Gasa crater, distinct gully channels extend up into many sub-alcoves. Cutoffs between these tributary channels and main trunk channels indicate multiple flow events (Fig. 10). Multiple terraces are common in the channels (e.g., Schon and Head, 2009) as well as streamlined islands or longitudinal bars (Fig. 11). Some lower

portions of the main channel regions (approximately 50 to 200 m in width) appear choked by sediment with discontinuous channel and bar features similar to braided stream environments (Fig. 12). The terrestrial conditions that give rise to braided stream morphologies, including high gradient, abundant sediment, and variable discharges (e.g., Leopold et al., 1964; Ore, 1964; Miall, 1978), are all likely to be applicable to the Gasa crater gullies. The sediment observed in the gullies appears uniform at HiRISE resolution (no lags or sorting is observable), however meter-scale boulders are observed in the channels and on alcove slopes (including boulders with associated boulder tracks from recent downslope movement). Analysis of gully apex slopes (the gradient where deposition begins) in Gasa crater by Kolb et al. (2010) found ten gullies with apex slopes consistent with “wet or fluidized emplacement” (16.3°-20.4°) and eleven gullies with steeper apex slopes (20.7°-26.4°) consistent with dry granular flows. Amongst their five study crater locations, the gullies in Gasa crater were best preserved (Kolb et al., 2010), which is consistent with the youthful age of Gasa (Fig. 2). They concluded that gullies are likely to have formed in the geologically recent past via a wet/fluidization mechanism and that subsequent and any present activity is likely “dry post-gully modification” (Kolb et al., 2010).

Gully Fans: Below the rocky outcrops of the alcove walls, many Gasa crater gullies exhibit sharply defined sedimentary interfluves (elevated ridges between channels of the same drainage network) (Fig. 8A and Fig. 12).

These sediment masses are bounded by gully channel and fan deposits on their margins. Boulders are observed on their sides, but bedrock is not outcropping (Fig. 12). Most are located directly downslope (in the lee) of bedrock outcrops and have sharp ridges. These are evidence of the abundant sediment available for erosion and transport.

Adjacent to the sedimentary interfluvies and extending downslope to the floor of the crater are the gully depositional fans. The fans have coalesced to form a large continuous sediment mass, similar to a bajada (a broad depositional deposit formed by the coalescing of alluvial fans) or pediment (veneer of eroded materials at the base of a mountain formed by scarp retreat) (Fig. 8A), though individual fans and depositional lobes are still distinguishable (e.g., Fig. 11). Both small and large channel diversions are present on the fans. Discontinuous channel segments are also observed, notably on lower portions of the fans. Channels dissecting fan surfaces are predominately linear, though some channels, especially on eastern fans, display modest sinuosity (Fig. 8A, Fig. 11A). Although the slopes are steeper, these characteristics are comparable to a terrestrial waterlaid alluvial fan (e.g., Blair, 2002).

The toes of the fans impinge on multiple crater floor textures. In the northwestern floor of the crater, gully fans extend onto rocky crater floor material that we interpret as slump deposits (Fig. 8A). Along the northern margin, fan material extends onto a hummocky-textured floor morphology.

Similar floor morphology was described by Tornabene et al. (2006) in other very young craters and has been interpreted as volatile-rich suevite that degassed rapidly in the terminal phases of the impact event (Boyce et al., 2011).

If Gasa crater and these gullies postdate the latitude-dependent mantle that is the source of gullies elsewhere in the area (Schon et al., 2009; Schon and Head, 2011), what is the source of meltwater that might have produced the gullies inside Gasa crater?

Evidence of Glacial Ice in the Host Crater

The 18-km diameter host crater within which Gasa crater formed contains latitude-dependent mantling deposits, hosts many gullies, and has a significantly lower depth:diameter ratio (0.07) than Gasa crater (0.12). In this section we outline multiple lines of evidence that this crater hosted a significant glacial deposit that predated the Gasa crater impact. The large gullies within Gasa formed in association with these ice deposits. The Gasa crater impact occurred into this ice-rich glacial substrate, common at these latitudes (Head et al., 2008; Dickson et al., 2011). Additional geological evidence for this scenario is provided by the action of meltwater and fluidized ejecta created by the Gasa crater impact that flowed on the host crater floor and ponded in local topographic lows (Fig. 13A).

Immediately north and northeast of the Gasa crater rim crest on the floor of the host crater ponded materials and associated channels are observed (Fig. 13). The ponds are sourced by channels that extend several kilometers from the lower crater wall (Fig. 13B). The channels are diverted around local topographic obstacles (Fig. 13B) and the ponds are observed where material pooled in local topographic lows (Fig. 13A). This topographic control of flow direction and evidence of flow back toward the Gasa crater rim indicate that these deposits are not the result of surging ejecta from the Gasa impact. The ponded material has a texture that is smooth at HiRISE resolution with surface fractures that we interpret as resulting from desiccation and contraction (Fig. 13C). The proximal channel feeding the pond (Fig. 13C) is composed of a central trough with leveed margins that we interpret as suggesting emplacement of the material by a wet debris flow.

We interpret these features to be the result of substantial melting of pre-existing debris-covered glacial deposits (Berman et al., 2005; Head et al., 2008; Dickson et al., 2011) induced by the Gasa impact and ejecta emplacement on the wall and floor of the host crater. These materials formed an unstable slurry within the northern portion of the host crater, which immediately flowed downslope and settled in topographic lows, moving generally downslope toward Gasa crater, but apparently not breaching the topographic rim crest (Fig. 13). Subsequent desiccation of the slurry led to the formation of contraction cracks. These features provide substantial evidence

that Gasa crater impacted partly into an ice-rich substrate formed by glacial deposits concentrated on the floor at the base of the pole-facing host crater wall (e.g., Head et al., 2008; Berman et al., 2009; Dickson et al., 2011).

Neither gullies nor similar ponded materials are found in the southern portion of the host crater (south of Gasa crater; Figs. 1, 8). Following the Gasa impact, melting of exposed debris-covered glacial ice in the Gasa crater wall would provide a source of water for gully activity within Gasa crater itself. The northern position of the glacial ice within the host crater (Fig. 13A) is consistent with the occurrence of gullies only in the northern walls of Gasa (Fig. 8).

Glacial Accumulations in Craters

Amazonian ice accumulations in craters have been interpreted from the presence of concentric crater fills (Levy et al., 2009c; Head et al., 2008; Dickson et al., 2011) as well as arcuate ridges interpreted as glacial moraines (Howard, 2003; Milliken et al., 2003; Arfstrom and Hartmann, 2005; Berman et al., 2005; Head et al., 2008; Berman et al., 2009). While concentric crater fill may suggest relatively homogenous ice distribution, moraines indicate preferential ice accumulation on crater walls and glacial flow (e.g., Benn and Evans, 1998; Head et al., 2008; Dickson et al., 2011). Like gullies (Balme et al., 2006; Dickson et al., 2007) and viscous flow features (Milliken et al., 2003), these moraines and associated spatulate depressions are found

preferentially with pole-facing orientations in the southern hemisphere (Berman et al., 2005; Head et al., 2008). Interpretations of crater densities led Arfstrom and Hartmann (2005) to suggest that these features are no more than 10 Myr old and are most likely to have formed during a high obliquity phase that ended approximately 4 Ma. Subsequent deformation, flow, and sublimation are responsible for more recent surface alteration and possible removal of small craters (Arfstrom and Hartmann, 2005).

Berman et al. (2009) specifically investigated the eastern Hellas region, including Promethei Terra. In that study they described commonly observed morphologies with pole-facing orientations indicative of ice accumulation and flow including lobate flows and arcuate ridges. Lobes were found to range in length from 1 km to 7 km with a mean length of 4.5 km (Berman et al., 2009). Arcuate ridges are found below gully alcoves and have similar widths (e.g., Berman et al., 2009). This spatial association has been interpreted as consistent with glaciers emanating from alcove cirques associated with gully formation during glacial retreat (Arfstrom and Hartmann, 2005; Berman et al., 2005; Head et al., 2008; Berman et al., 2009). In a stratigraphic analysis of recent crater modification, Head et al. (2008) showed that gullies are younger and occur after such glacial deposits have lost ice. In these settings, ice preferentially accumulates on pole-facing walls sufficient to initiate ice flow and the development of debris covered glaciers. As climate conditions changed due to a general decrease in obliquity and the

period of glaciation waned, these glacial systems lost ice via sublimation, particularly in their accumulation zones, exposing hollows and forming spatulate depressions (Head et al., 2008). Gullies are observed stratigraphically to post-date the loss of glacial ice; they occur in the regions that would have been accumulation zones for the glacial ice. Therefore, Head et al. (2008) suggest a genetic relationship based on ice accumulation in the shared source regions for the debris-covered glaciers and the later gullies. Gully activity led to the deposition of fans in the spatulate depressions. While stratigraphically older fans are deformed, younger fans are un-deformed (Head et al., 2008).

Consistent with these interpretations of glacial systems, the region east of Hellas is also identified in global climate modeling studies as a location of enhanced ice accumulation under past obliquity conditions (Forget et al., 2006; Madeleine et al., 2009) and contains many lobate debris aprons (Pierce and Crown, 2003). In craters neighboring Gasa crater, evidence of preferential glacial ice accumulation and flow is pervasive. Detailed observations from one of these craters provide additional evidence that a substantial glacial ice accumulation was present in the host crater at the time of the Gasa crater impact. Approximately 70 km north of Gasa crater is a 13.5-km diameter crater (Fig. 14). This crater has a mottled and lineated fill material similar to those described by Kreslavsky and Head (2006) in a study of glacially modified craters in the northern high-latitudes. Spatulate

depressions bounded by moraine-like ridges are associated with the largest gully alcoves (Fig. 14). Linear ridges are parallel to the base of the pole-facing crater wall. We interpret these features to be the result of glacial accumulations concentrated on the pole-facing wall. Because gully fans deposit into the spatulate depressions, glaciation preceded the most recent gully fan deposition here (Fig. 14), which is consistent with chronological interpretations in other studies (Arfstrom and Hartmann, 2005; Head et al., 2008; Berman et al., 2009).

Comparison with Fresh Crater Zumba

To test the hypothesis that the gullies in Gasa crater result from impact into an ice-rich substrate we investigated the population of rayed craters identified by Tornabene et al. (2006) for a comparable impact. None of these craters contain gullies. While Gratteri (D = 6.9 km) and Tomini (D = 7.4 km) are most similar to Gasa crater in diameter, they are located at more equatorial latitudes where gullies are not found. Zumba crater (28.65S, 226.9E) is the most pole-ward rayed crater identified by Tornabene et al. (2006) and is located at a latitude near to where gullies are common. Zumba occurs in Daedalia Planum on a known substrate of Hesperian age lavas (Scott and Tanaka, 1986).

The rim of 2.6-km diameter Zumba crater is un-crenulated (Fig. 15A). Uniform debris chutes line the rim regardless of orientation. The crater wall

has a uniform smooth texture of unconsolidated materials with no evidence of incision (Fig. 15A). Slumped materials are deposited on the crater floor. An extensive crater count of 46-km² of the ejecta deposit revealed 1197 superposed craters (Fig. 15B). Isochrons of Hartmann (2005) imply a best-fit age of 0.8 Ma for Zumba crater, consistent with a crater retention age of 0.2 to 0.8 Ma reported by Hartmann et al. (2010). Therefore, Zumba is similar in age to Gasa crater, but has very different morphology due to its impact into a uniform substrate (Fig. 15A) rather than the glaciated crater interior environment that Gasa crater impacted (Fig. 1 and Fig. 13).

Chronological Interpretation

Originally disparate individual observations can be assembled into a chronology (Fig. 16) that documents the history of gullies and the transitions in their activity. Previously, gullies were well known on older planetary surfaces, e.g., Dao Valles (Bleamaster and Crown, 2005) as well as in craters (e.g., Heldman and Mellon, 2004; Balme et al., 2006; Dickson et al., 2007), but Gasa crater presents a compelling example of gullies also developing in an un-ambiguously very young location (Fig. 2). Chronological constraints require gully formation in the recent glacial/interglacial epoch (Schon et al., 2009a) and indicate that gullies can develop solely from degradation and melting of ice-age mantling deposits (Schon and Head, 2011).

The Eastern Promethei Terra region provides an opportunity to synthesize observations of multiple gullies, their geologic settings, and several temporal constraints to more fully understand the occurrence, ages, and formation processes of mid-latitude gullies. Principal geological events of interest (Fig. 16) were preceded by formation of the Noachian cratered terrain. Larger gullies of the Gasa host crater predate the Gasa impact and the emplacement of latitude dependent mantling deposits (Fig. 5). The glacial systems interpreted to be responsible for features such as arcuate and moraine-like ridges are most likely to date from a phase of higher mean obliquity, $\sim 10^{-4}$ Ma (Arfstrom and Hartmann, 2005), though older glacial episodes elsewhere are also known (e.g., Head et al., 2005). Debris flow deposits (Fig. 13) induced by the Gasa impact provide geological evidence that the pole-facing crater wall and floor region of the host crater contained relict glacial deposits that pre-dated Gasa (Fig. 16).

Emplacement of hemispheric-scale ice-rich mantling deposits during the previous ice age (2.1 – 0.4 Ma) was cyclic and latitude-dependent (Fig. 3; Head et al., 2003). The youthfulness of high latitude mantling is known from the pervasiveness of un-modified polygonally patterned ground (Kreslavsky et al., 2011). Mid-latitude mantle surfaces are older and show more evidence of degradation than higher latitude mantling which was likely to have been emplaced more recently (Kostama et al., 2006). Secondary craters from the Gasa crater impact are superposed on the mantle surface (Fig. 4) and no

mantling is observed within Gasa (e.g., Fig. 8). These observations require that the Gasa impact occurred subsequent to the most recent mantle deposition in this region. The size-frequency distribution of superposed craters (Fig. 2) suggests that the Gasa impact occurred between 2.4 Ma and 0.6 Ma with a best fit to Hartmann (2005) isochrons of 1.25 Ma. Together, the crater retention age and superposition relationship (secondary craters on the mantle; Fig. 4) indicate that the Gasa impact occurred during the waning of the most recent ice age on Mars.

Gullies are known to be active after the Gasa crater impact from gully fan deposition over Gasa secondary craters (Fig. 7; Schon et al., 2009a; Schon and Head, 2011) and the two generations of gullies in the host crater (Fig. 6). This is consistent with gully formation resulting from melting and degradation of latitude-dependent mantling deposits (Schon and Head, 2011). Thus, there were two ice-rich deposits in this region at the time of the Gasa impact, 1) regional near-surface ice deposits from the ice-age latitude-dependent mantle (e.g., Head et al., 2003), and 2) remnant glacial ice deposits in the interiors of some impact craters, such as the host crater and others (Figs. 1, 13, 14) (e.g., Berman et al., 2005; Head et al., 2008; Berman et al., 2009; Dickson et al., 2011). Gully formation within the Gasa crater interior resulted from the impact disruption of these glacial ice deposits and their exposure in the pole-facing crater wall. Thus, ice-rich erodible substrate for gully activity was made available by the Gasa impact into the host crater

interior target material. Abundant immediately-post-Gasa debris flow features (Fig. 13) provide additional evidence of the presence of ice on the crater floor that was melted and exposed by the impact. Gully evolution in Gasa was likely to have been limited by the ultimate quantity of meltwater generated from the ice deposits. As this source of meltwater waned, steep slopes and abundant sediment remained available for dry mass wasting processes, consistent with our observations of fresh boulder tracks and the apex slope analysis of Kolb et al. (2010).

Conclusions

Observations show that meltwater from the ice-rich LDM is the most common source of liquid for erosion and transport in gully systems. However, formation of gullies in the late glacial period-aged Gasa crater occurred due to impact into a debris-covered glacial substrate that provided an additional source of meltwater for gullying. Gasa crater is the most poleward (35.7°S) rayed crater identified to date on Mars, approximately seven degrees further south than Zumba (28.7°S), the most poleward rayed crater identified in the survey of Tornabene et al. (2006). Gasa crater has a crater retention age of ~ 1.25 Ma. The superposition of Gasa crater secondary crater chains on latitude-dependent mantling deposits in the region and the lack of mantling within Gasa crater show that Gasa crater post-dates the last episode of visible mantle deposition at this location. Because deposition of mantling

deposits is interpreted to be coincident with obliquity excursions during Mars most recent ice age (Head et al., 2003), these relationships confirm the youthful age of Gasa crater.

Gasa crater proximal and distal stratigraphic relationships show that:

1) gully activity extends to extremely young ages, occurring at least as recently as 2.1-0.4 Ma, and perhaps even more recently, 2) gully activity is favored on steep, pole-facing slopes and involved multiple stages of development, 3) gully activity is very closely associated with erosion and degradation of the layered ice-rich LDM, 4) gullies commonly source in the ice-rich layers of the LDM, 5) close association of gullies with the ice-rich LDM layers implicates liquid water in their formation, 6) Gasa crater formed just after the last emplacement of the LDM at these latitudes, 7) robust gullies formed in Gasa despite the lack of emplaced and exposed LDM layers, 8) the source of meltwater for the anomalous gullies in the interior of Gasa can be traced to the excavation and exposure of a debris-covered glacier on the floor of the parent crater in which Gasa formed. Thus, the Gasa crater example has provided new insights into the ages, geographic and stratigraphic associations, relation to recent ice ages, and sources of water and associated climate conditions that led to gully formation.

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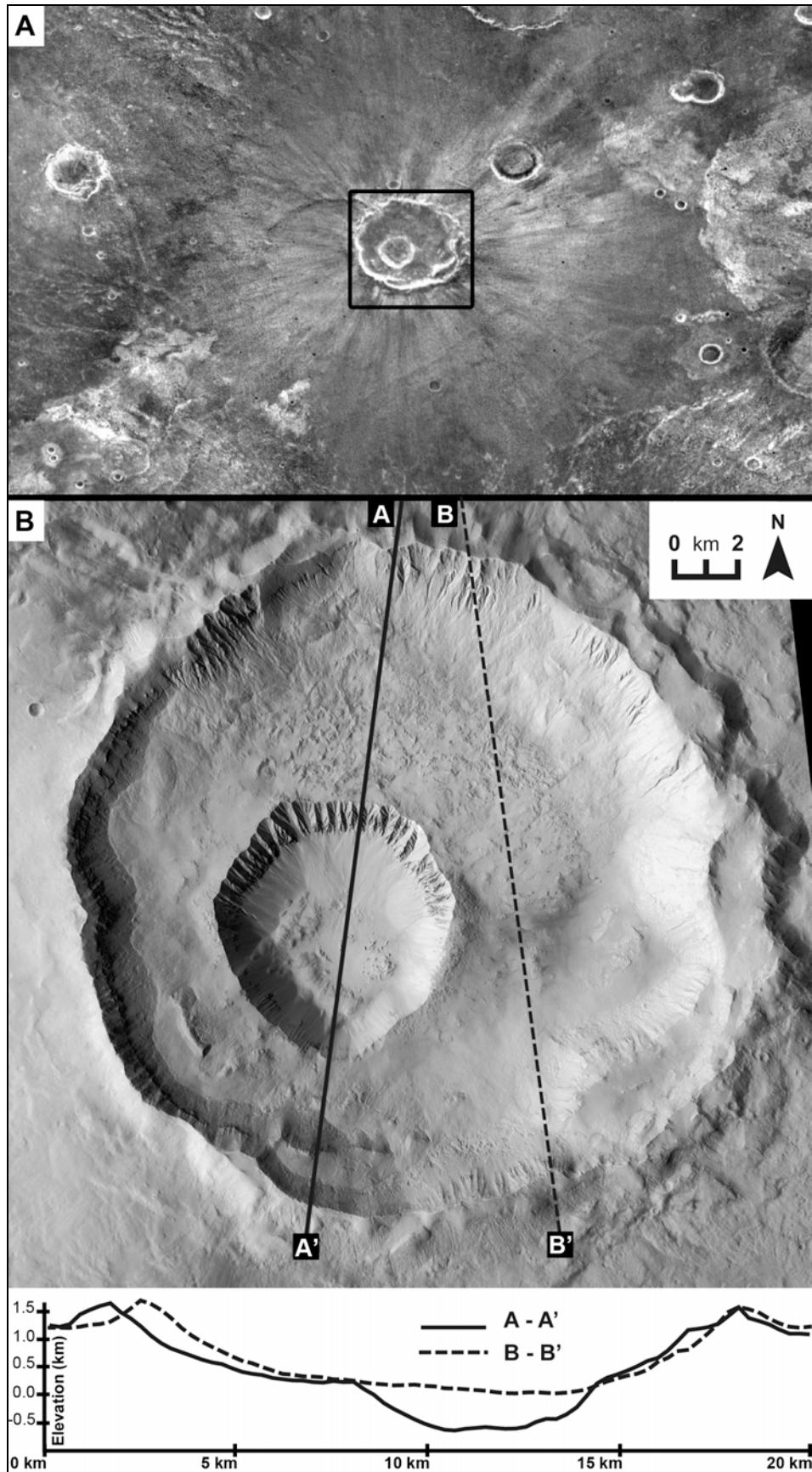


Figure 1: Context of Gasa crater. (A) THEMIS (Thermal Emission Imaging System) nighttime thermal infrared images show a prominent pattern of fresh rays that emanate from Gasa crater. (B) Gasa crater is offset on the floor of an un-named ~18-km diameter degraded crater. Portion of CTX: P08_004060_1440_XI_36S230W. Topographic profiles from the Mars Orbiter Laser Altimeter (MOLA) show the prominence of Gasa crater (A-A') within the host crater. Pole-facing slopes are shallower in both craters compared to equator-facing slopes. Profile B-B' of the host crater shows a gently poleward dipping crater interior that is characteristic of glacial deposits (Head et al., 2008).

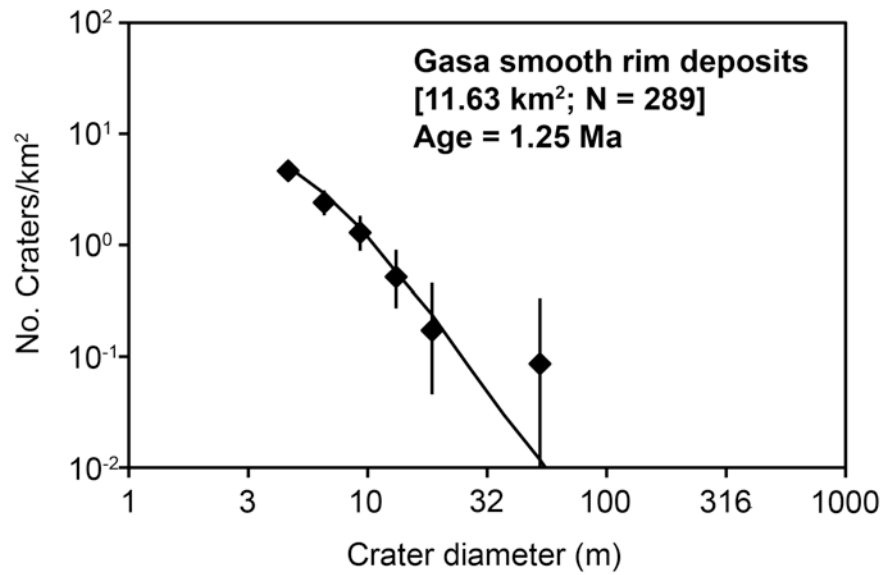


Figure 2: Gasa crater age. Crater count data of Schon et al. (2009a) on smooth near-rim deposits of Gasa crater using HiRISE data

(PSP_004060_1440) are shown on an incremental size-frequency plot.

Hartmann (2005) isochrons suggest a crater retention age of ~1.25 Ma for the Gasa crater impact event.

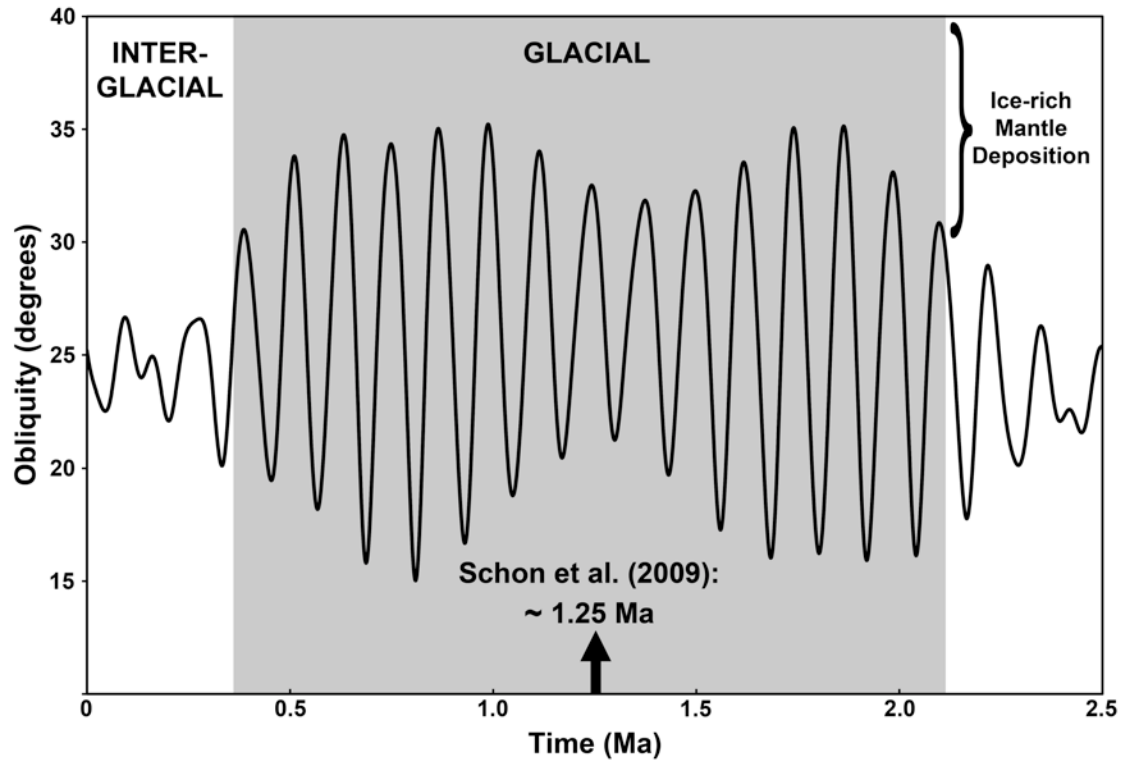


Figure 3: Mars obliquity record (Laskar et al., 2004). The deposition of multiple ice-rich mantling layers during a period of higher obliquity amplitude variations has been inferred from geological evidence (glacial period; Head et al., 2003). Meltwater derived from degradation of this mantling unit is interpreted to be responsible for gully formation (Dickson and Head, 2009). Crater retention age dating suggests that Gasa crater formed ~1.25 Ma, while the relative stratigraphic position of crater chains from Gasa crater on the mantle surface (Fig. 4) indicate that the development of widespread mantling deposits at this latitude had ceased by the time of the Gasa crater impact.



Figure 4: Gasa crater secondary crater chains (34.82°S, 128.30°E). These crater chains are located ~75 km northwest of Gasa crater on the regional latitude-dependent mantle surface and show no evidence of subsequent modification. Portion of CTX: P15_006895_1430_XN_37S231W.

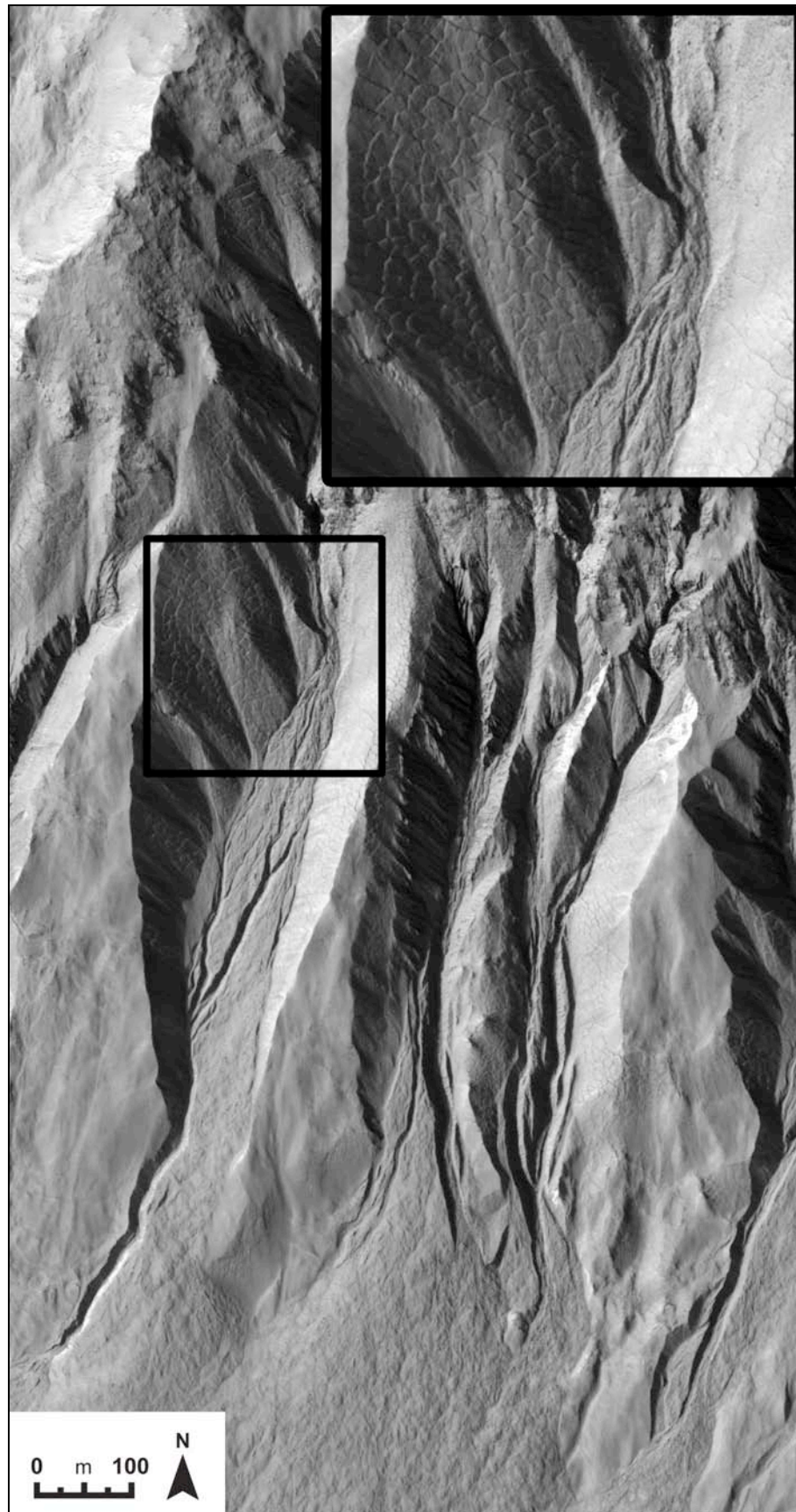


Figure 5: Host-crater gully alcove polygons. The host crater has several gullies that are characterized by polygonal patterning of the alcove walls. The formation of such polygons is interpreted as requiring the presence of an ice-rich substrate (Levy et al., 2009a), such as the latitude-dependent mantle (Head et al., 2003), and has previously been observed in association with other gullies (Levy et al., 2009b). The presence of inner channels and discontinuous channel segments suggest multiple flow events. Portion of HiRISE: PSP_005616_1440.

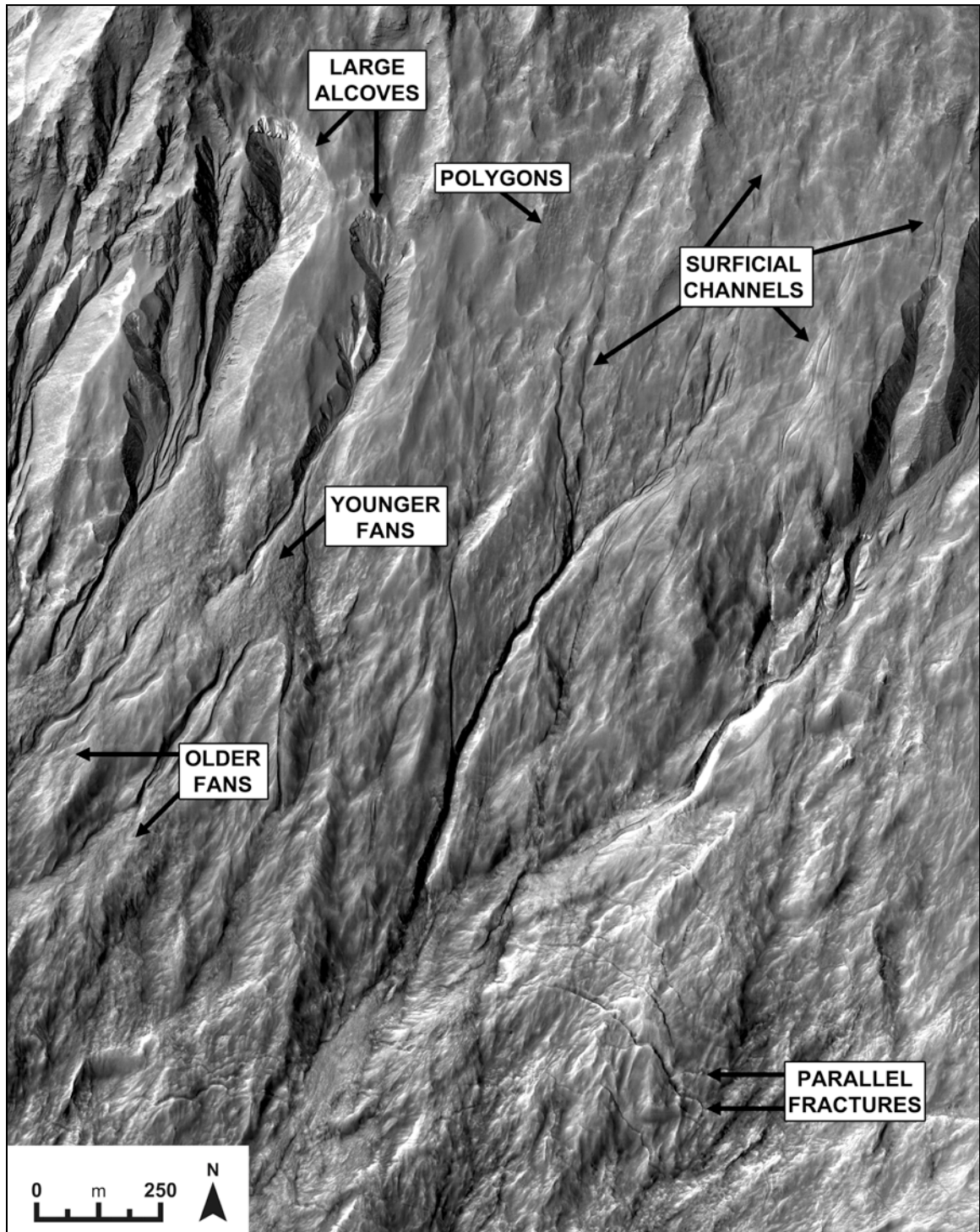


Figure 6: Host crater gullies. Older gullies fans associated with larger alcoves are modified by parallel fractures (e.g., Head et al., 2008). Younger fans are unmodified and overprint older gully channels. Polygonal terrain is

observed on the latitude-dependent mantle that blankets the crater wall. Alcove-less surficial channels emerging from the latitude-dependent mantle extend for hundreds of meters and feed into larger gully drainages. These stratigraphic relationships suggest multiple generations of gully activity. The muted and degraded topography of the older gullies suggests that they predate the Gasa cratering event. Portion of HiRISE: PSP_005616_1440.

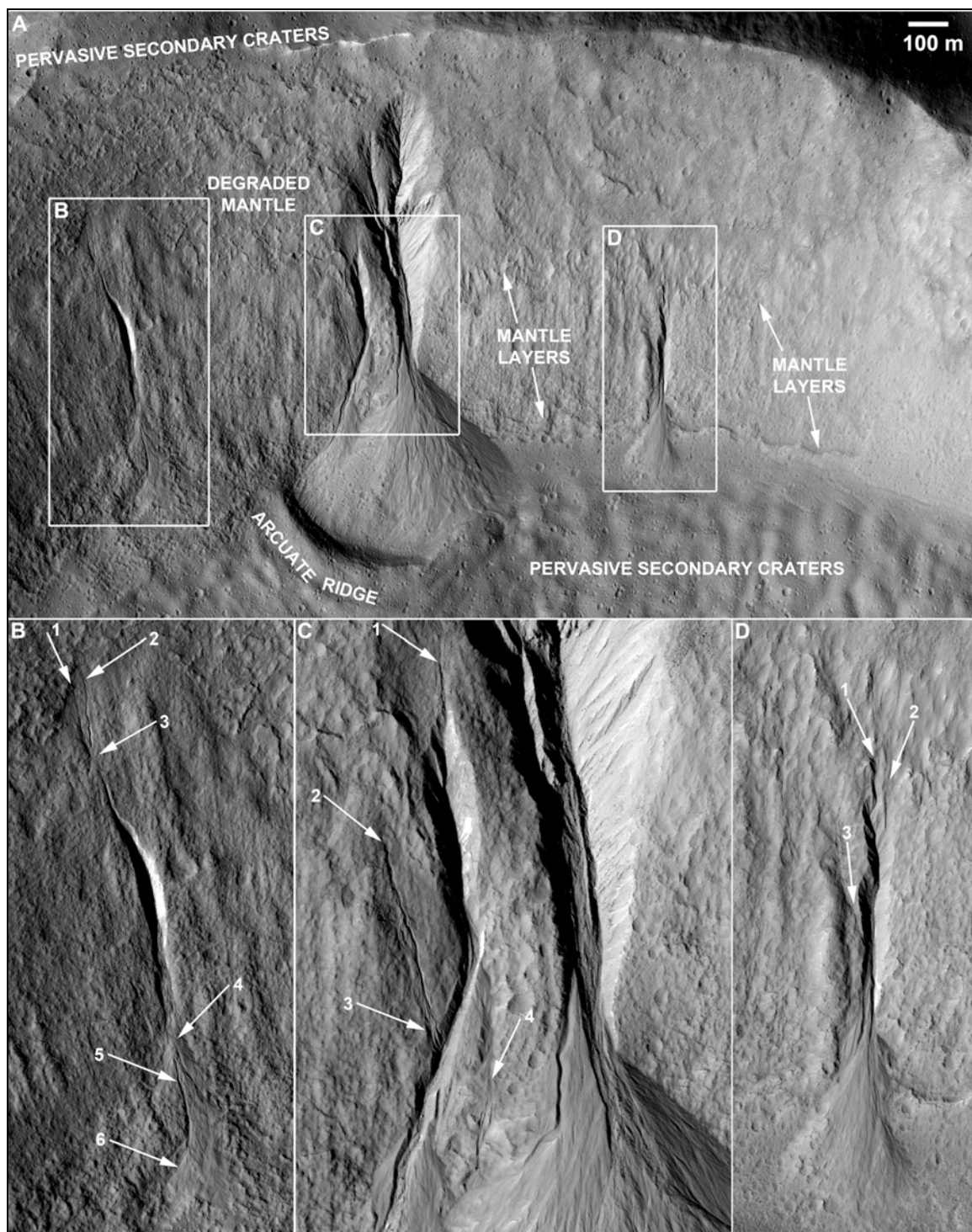


Figure 7: Approximately 100 km northeast of Gasa crater, gully fans in a 5-km crater (A) post-date a dense population of Gasa secondary craters (eastern Promethei Terra, $\sim 35^{\circ}\text{S}$, 131°E). The pole-facing wall (shown) is composed of

degraded layers of ice-rich latitude-dependent mantling deposits. On the crater wall, post-Gasa degradation of the mantling unit has led to a partial to nearly complete removal of the secondary crater population and exposed prominent layers of the mantle. Small channels and gullies without alcoves emerge from the degraded mantle and are interpreted to have deposited fans during degradation and melting of the ice-rich mantling material. These fine-scale surficial gully features are highlighted with arrows in insets (B–D). These observations implicate meltwater from the degradation of latitude-dependent mantling in the process of gully formation. Portions of HiRISE PSP_002293_1450; from Schon and Head, 2011.

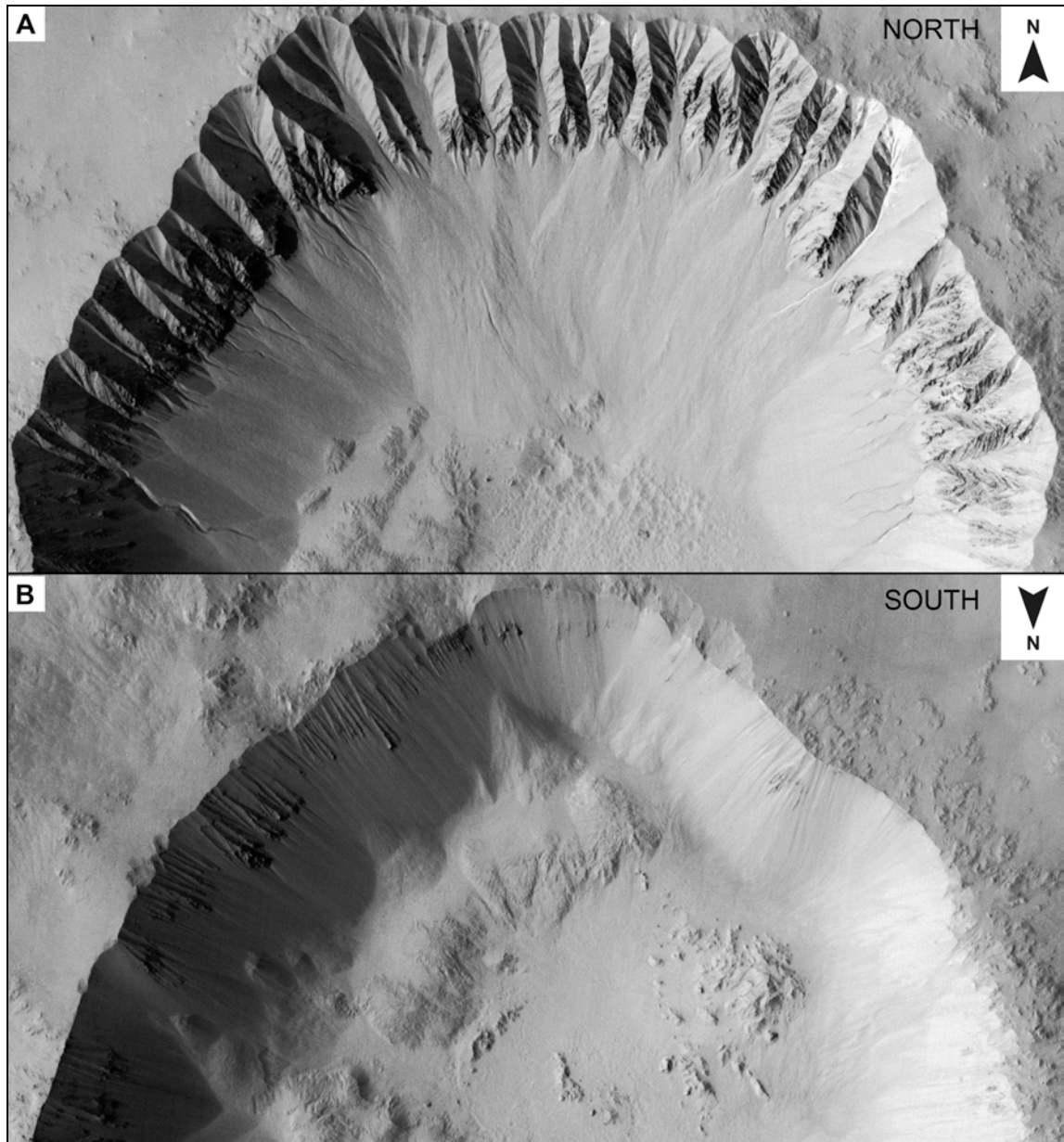


Figure 8: Asymmetry of Gasa crater rim crenulation and wall morphology.

(A) The northern half of Gasa crater has a crenulated rim due to the development of significant gully alcoves. These alcoves are most well-developed in the central portion of pole-facing wall and become progressively smaller at offset orientations. Gully fans have coalesced to form a large pediment. These fans extend onto a hummocky crater floor texture (bottom

center) that has been interpreted by Boyce et al. (2011) as resulting from rapid degassing of volatile-rich suevite immediately following the cratering event. Discontinuous channel segments are observed on the fans. (B) The southern half of Gasa crater has been flipped over the horizontal axis for ease of comparison (East is to the right in both panels). This portion of Gasa crater lacks gully alcoves and the rim is significantly less crenulated. In comparison to the gully fans and large sedimentary bajada developed in the north (A), talus cones and landslide deposits provide the only evidence of downslope movement on these equator-facing slopes. Portions of CTX: P08_004060_1440_XI_36S230W.

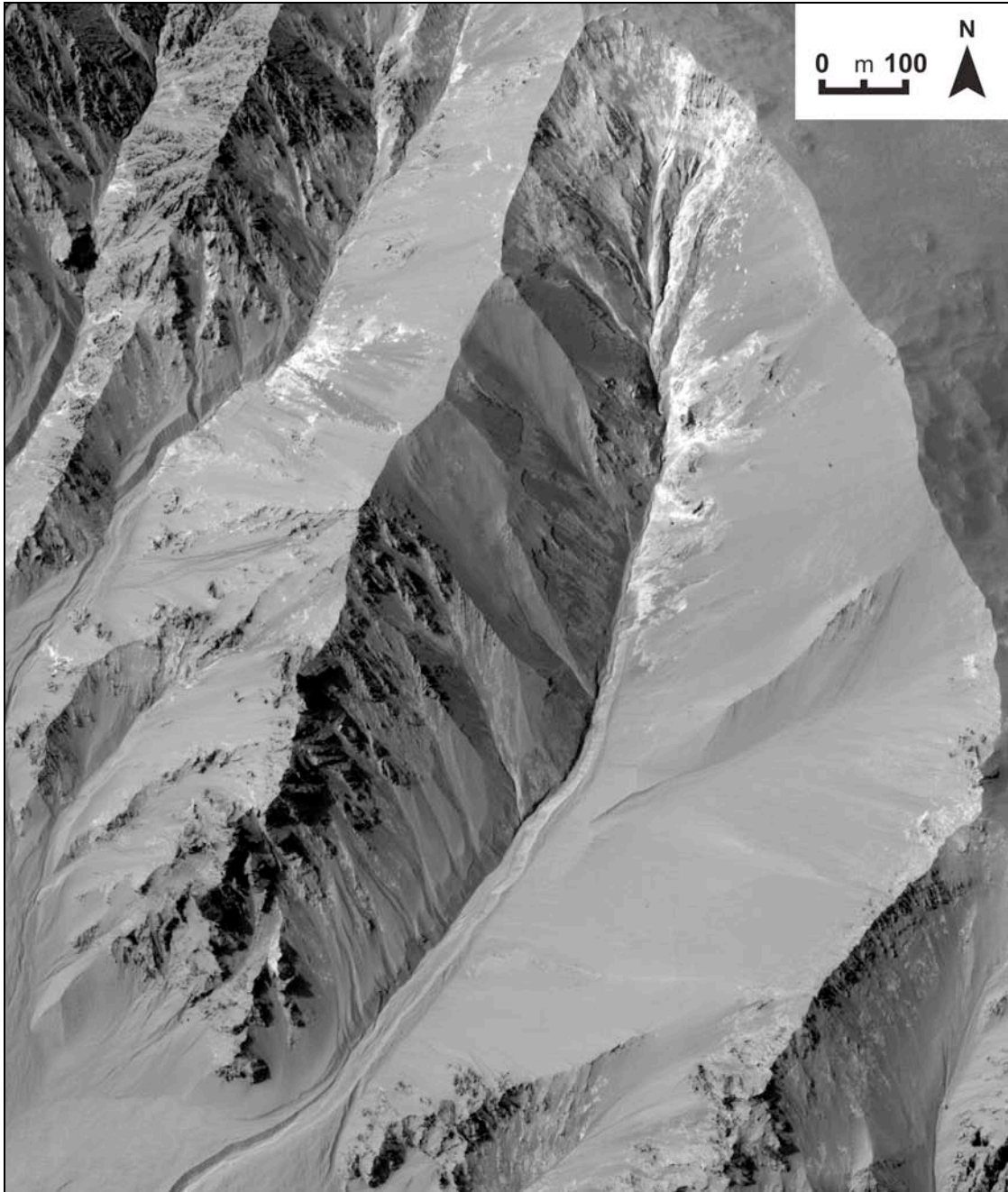


Figure 9: Gully alcove asymmetry in Gasa crater. Gully alcoves can also contain microenvironments based on orientation-dependent asymmetry (e.g., Williams et al., 2009; Morgan et al., 2010). In this example, only pole-facing gully alcove walls are incised by channels feeding the main gully channel. The preferential exposure of rocky material on the pole-facing walls also

suggest that these surfaces have undergone more extensive erosion than their equator-facing counterparts. Portion of HiRISE: PSP_005550_1440.

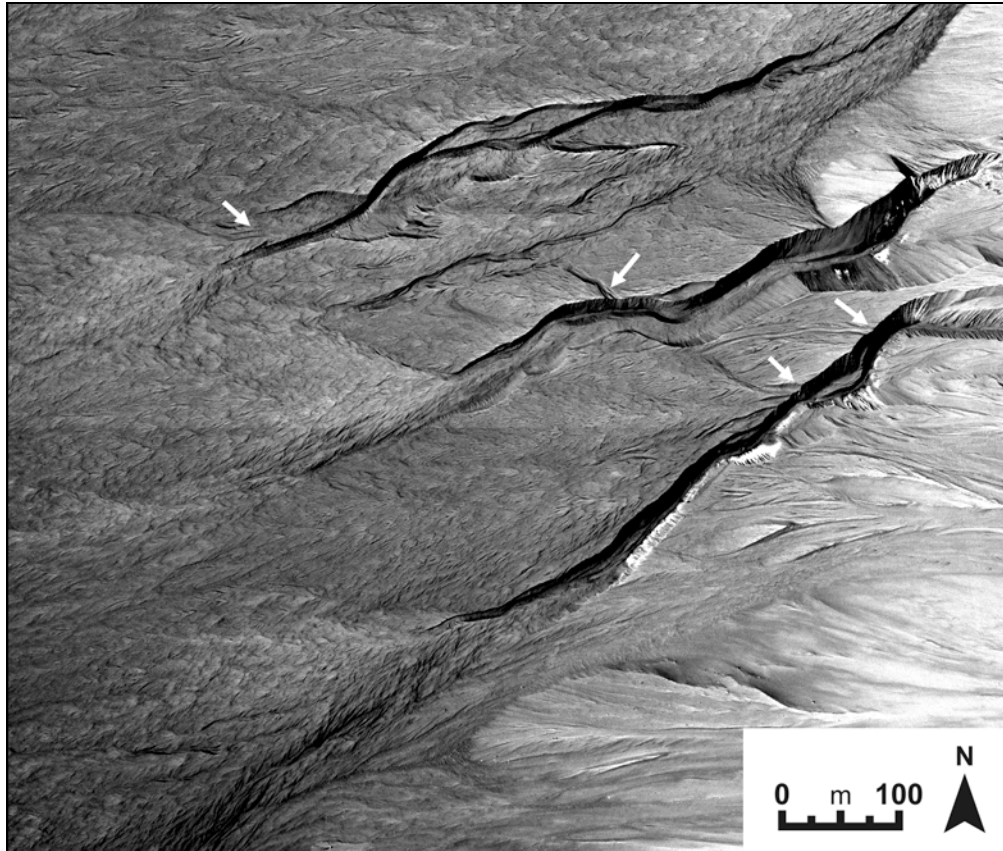


Figure 10: Example of gully channel incision and deposition in Gasa crater. Gully channels in Gasa crater contain many examples of channels that have been cut off by further channel erosion and diversion (arrows). Lighter-toned younger fan deposits are visible at left. Portion of HiRISE: PSP_004060_1440.

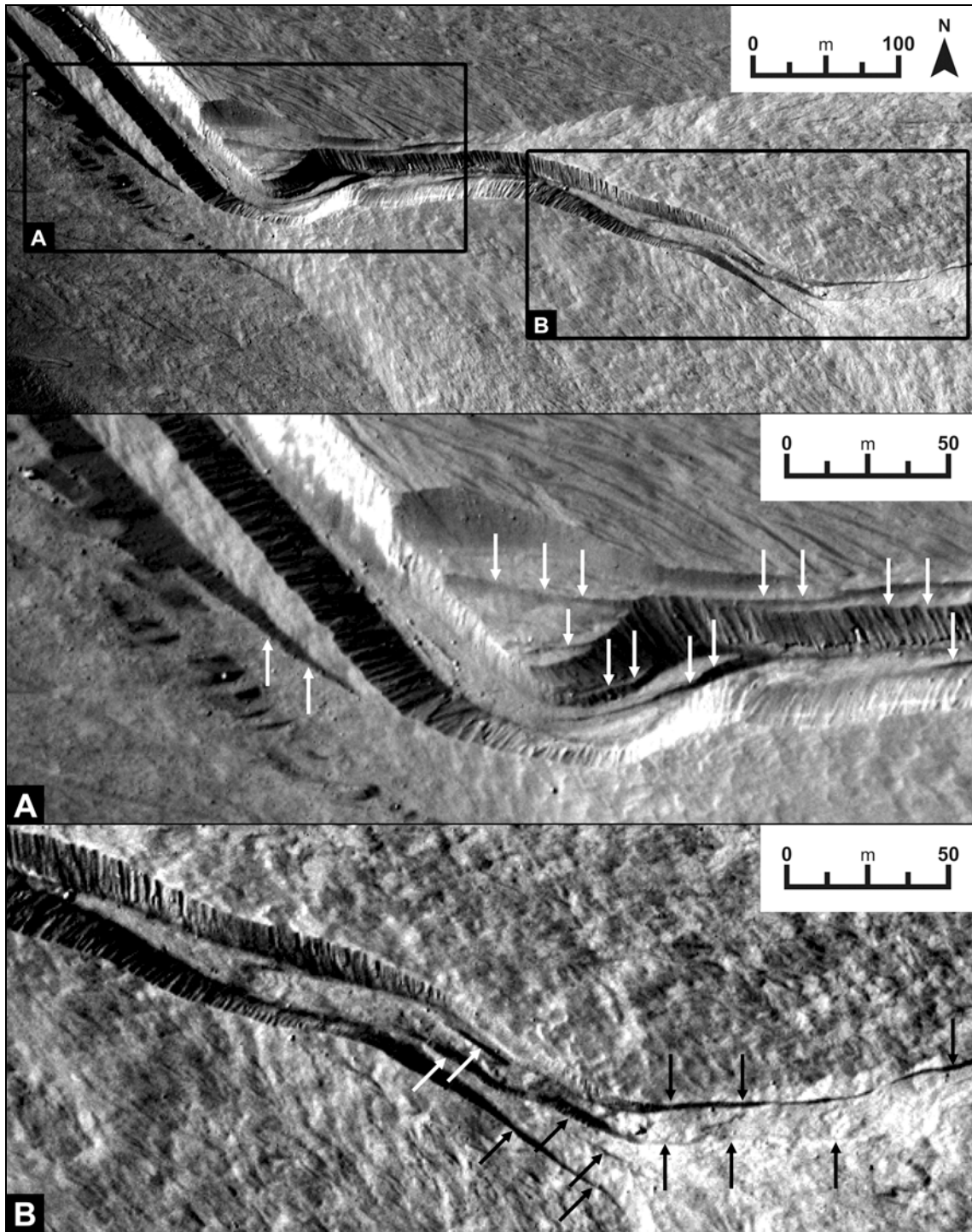


Figure 11: Example of gully fan deposition and terraces within gully channels in Gasa crater. Gully channels are observed eroding sediment previously transported and deposited by the gully. Terraces (arrows) are the

result of further channel incision and are commonly observed in terrestrial braided stream environments (Ore, 1964). Portion of HiRISE: ESP_014081_1440.



Figure 12: Braided stream morphology of Gasa crater gully channels.

Sedimentary interfluves are observed in the lee of alcove bedrock outcrops and are evidence of the abundant sediment available for erosion and transport. Individual sub-alcoves and channels feed into a larger sediment-choked channel region that exhibits morphology characteristic of terrestrial braided streams. Channel segments are cutoff, discontinuous, and choked by sediment in places indicating that later flow events may have been lower energy. Portion of HiRISE: ESP_014081_1440.

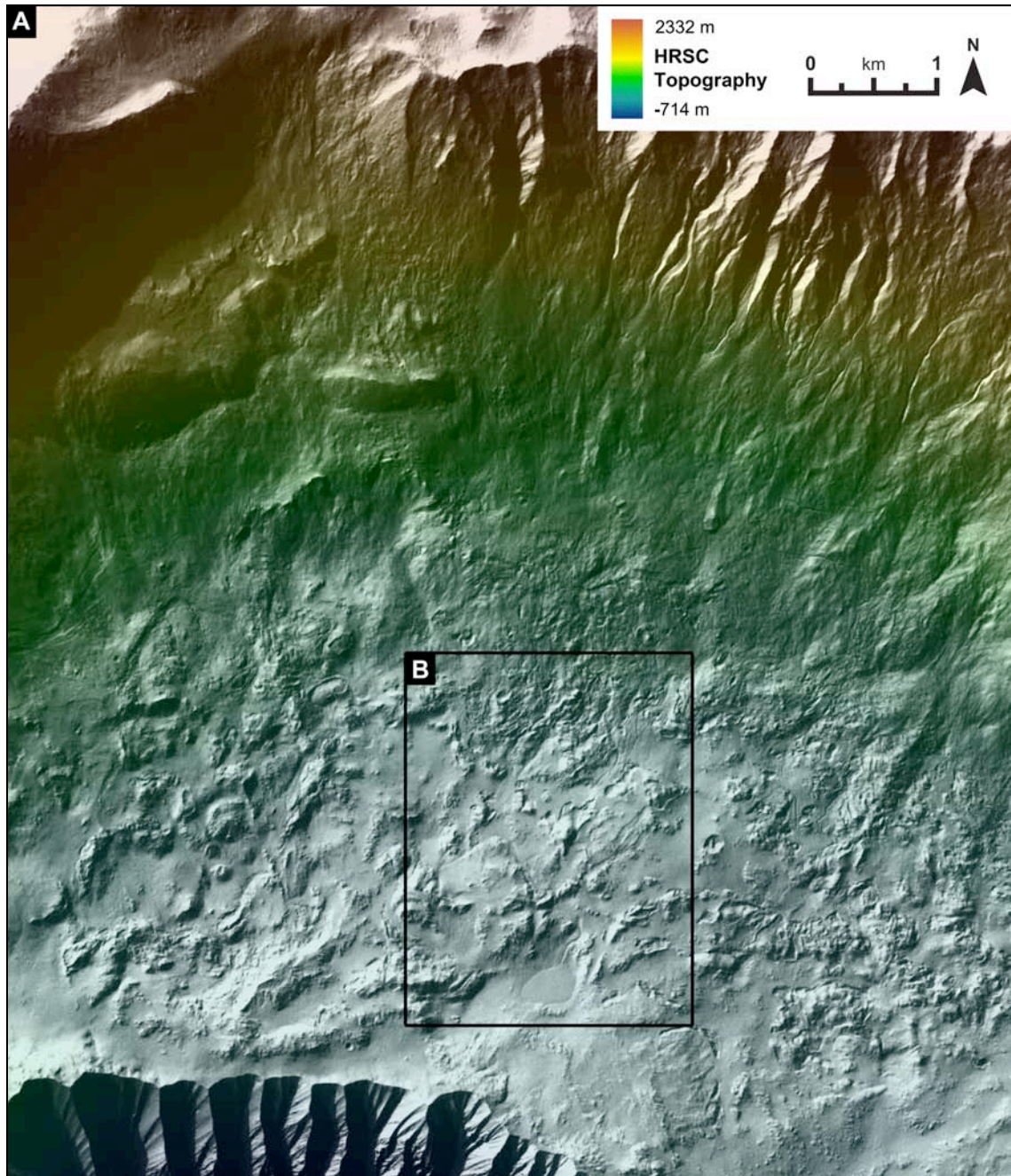


Figure 13A: Debris flow deposits on the host crater floor (Fig. 1). Ponded materials and associated channels indicate flow over a distance of approximately four kilometers. High-resolution topography data from HRSC and HiRISE digital terrain models show topographic control of the channels and ponded material indicating that these features are not the result of

surging ejecta. Portions of CTX: P16_007396_1453_XI_34S230W and HRSC:
h6494.

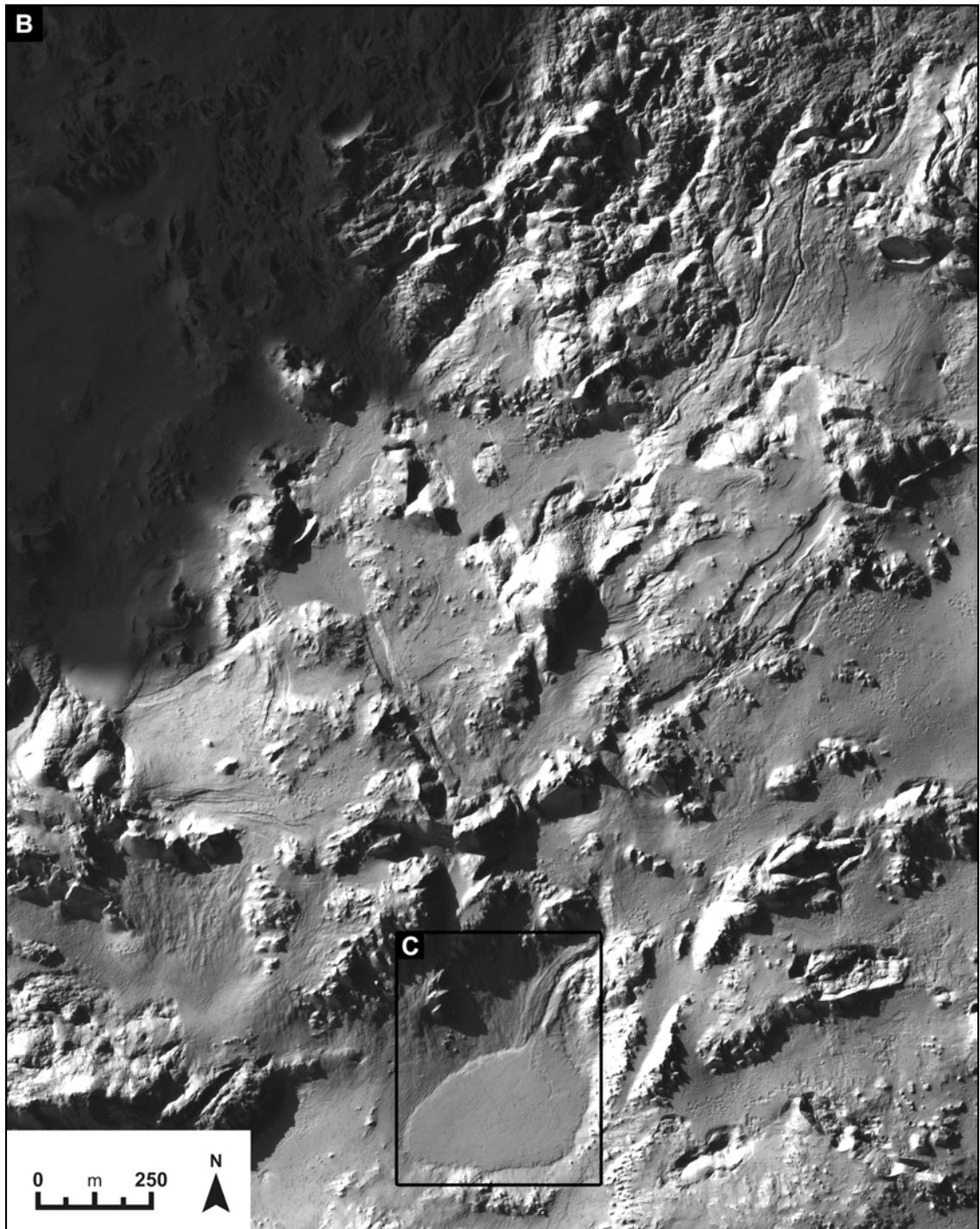


Figure 13B: Channels are observed from the lower wall of the crater to ponded material in a topographic low near the Gasa rim crest. We interpret this feature to be the result of the Gasa impact event depositing hot ejecta on top of a debris-covered glacier on the northern crater floor, and melting

glacial ice buried at shallow depth, forming a slurry that drained in directions related to local topography. Portion of HiRISE: PSP_009901_1440.

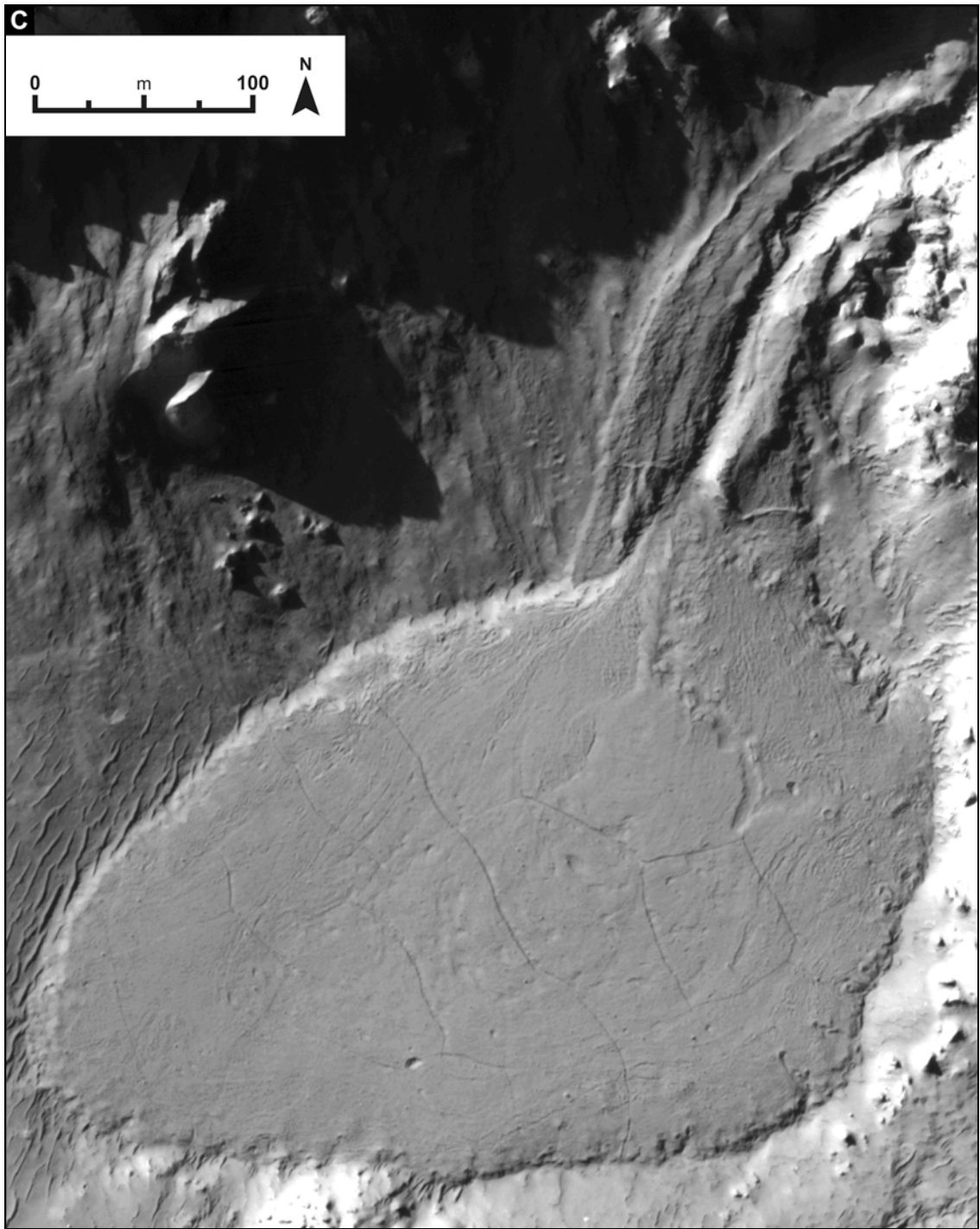


Figure 13C: Ponded debris flow material exhibits surface fractures interpreted as resulting from desiccation and contraction. The texture is homogenous and smooth at HiRISE resolution. Portion of HiRISE: PSP_009901_1440.

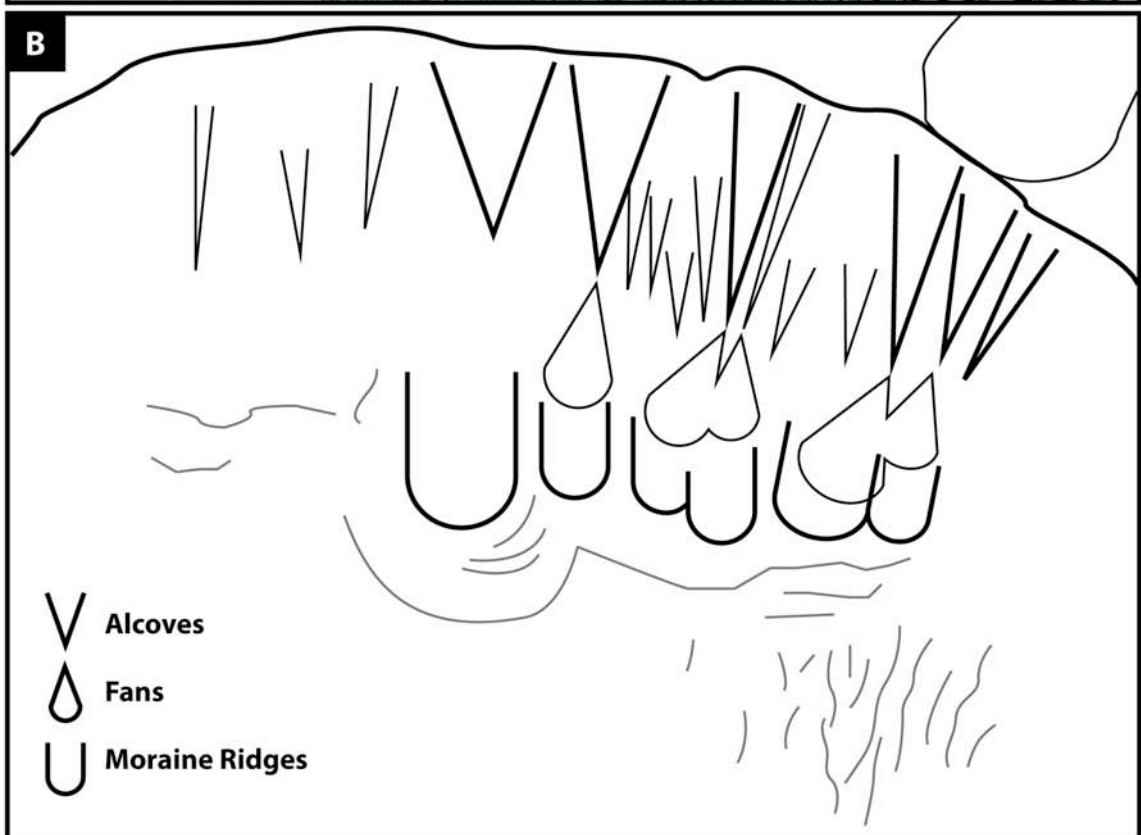
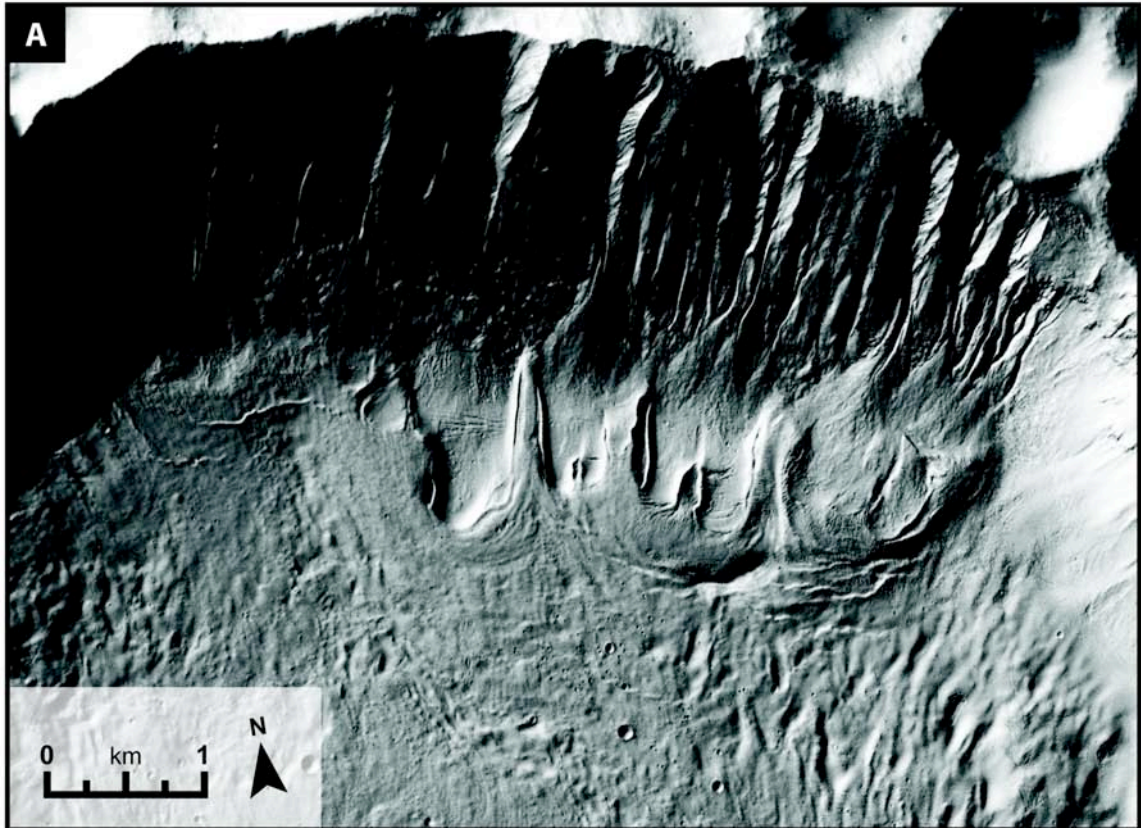


Figure 14: Nearby crater (~70 km north of Gasa crater; 34.45°S, 129.15°E) with evidence of debris-covered glacier deposits, and younger gully activity in the spatulate depressions of the older glacial deposits (see Head et al., 2008). This 13.5-km diameter crater is slightly smaller than the host crater and located further equatorward, but still contains prominent geomorphic evidence of past glacial ice accumulation and flow in the form of arcuate ridges and spatulate depressions. This type of glacial episode preceded the most recent gully activity in the crater. Portion of CTX: P16_007396_1453_XI_34S230W.

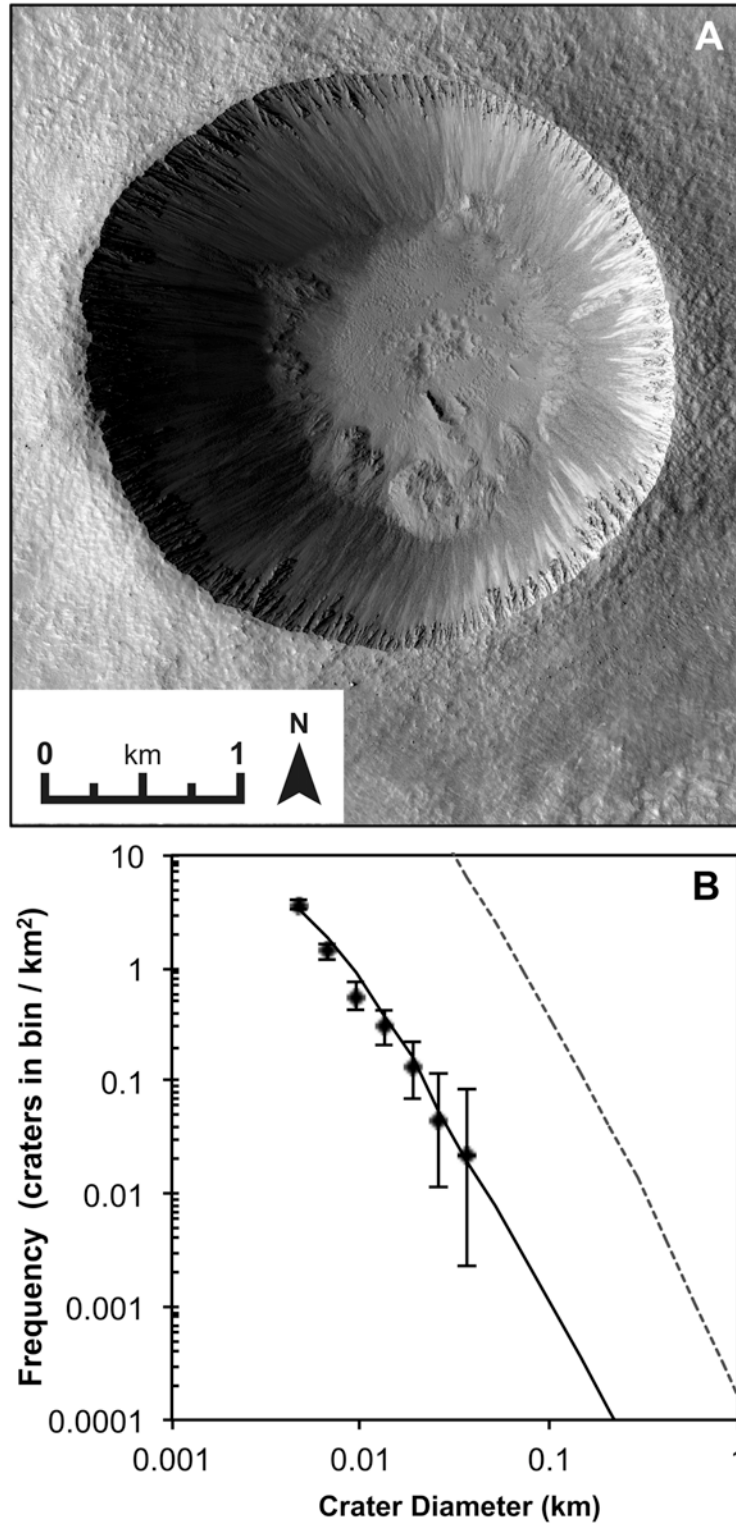


Figure 15: (A) Zumba Crater (28.7°S, 226.9°E). Zumba is the most poleward rayed crater (D = 2.6 km) identified by Tornabene et al. (2006) and is

therefore useful as a morphological comparison to Gasa crater. Zumba crater has a uniform un-crenulated rim, no gullies, and no evidence of crater wall incision. Portion of HiRISE: PSP_003608_1510. (B) Crater counting on the Zumba ejecta deposit revealed 1197 craters. Isochrons of Hartmann (2005) imply a best-fit age of 0.8 Ma, consistent with the age range of 0.2 Ma to 0.8 Ma reported by Hartmann et al. (2010) for the age of Zumba crater. Therefore, Zumba crater and Gasa crater are comparable in age, but while Gasa impacted a unique ice-rich crater interior, Zumba impacted into a uniform volcanic terrain.

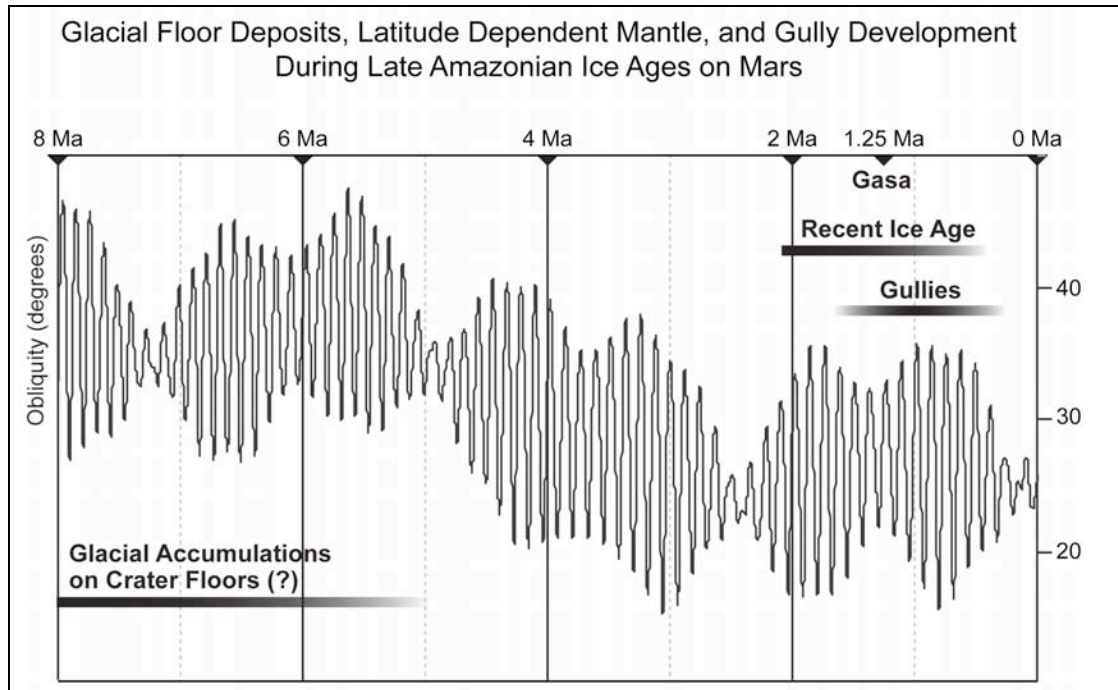


Figure 16: Timeline of glacial and gully activity during the Latest Amazonian period of Mars history. Mantling deposits are interpreted to have been emplaced in a latitude-dependent manner during the previous period of enhanced obliquity (Head et al., 2003). The obliquity data are from Laskar et al. (2004). The Gasa impact event is constrained by its crater retention age (Fig. 2) and the relative stratigraphic position on the latitude-dependent mantle (Fig. 4). Timing of gully activity is known to stratigraphically post-date the Gasa impact event in this area (Schon et al., 2009). Crater-wall glacial accumulations are suggested to date from a period of higher mean obliquity that ended approximately 5 Ma (Arfstrom and Hartmann, 2005).

CHAPTER 5

DECAMETER-SCALE PEDESTAL CRATERS IN THE TROPICS OF MARS: EVIDENCE FOR THE RECENT PRESENCE OF VERY YOUNG REGIONAL ICE DEPOSITS IN THARSIS.

(Submitted as Schon, S.C., and J.W Head (2011) Decameter-Scale Pedestal Craters in the Tropics of Mars: Evidence for the Recent Presence of Very Young Regional Ice Deposits in Tharsis, *Earth and Planetary Science Letters*, revised.)

Abstract

Global climate models predict that ice will be deposited in tropical regions during obliquity excursions from the current mean obliquity of $\sim 25^\circ$ to $\sim 35^\circ$, but no geological evidence for such deposits has been reported. We document the presence of very small (decameter scale) pedestal craters in the tropics of Mars (the Daedalia Planum-Tharsis region) that are superposed on an impact crater dated to ~ 12.5 million years ago. The characteristics, abundance, and distribution of these small pedestal craters provide geological evidence that meters-thick ice accumulations existed in the tropical Tharsis region of Mars in the last few million years when mean obliquity was $\sim 35^\circ$ (~ 5 - 15 Ma) before it transitioned to a mean of $\sim 25^\circ$ (~ 0 - 3 Ma). A

reconnaissance survey reveals similar small pedestal crater examples superposed on the older Amazonian Arsia Mons tropical mountain glacier deposit, suggesting that ice can accumulate in these tropical regions without initiating large-scale glacial conditions. These results support the predictions of general circulation models that ice can migrate to the equatorial regions during periods of moderate obliquity and then serve as a source for mid-latitude deposits.

Introduction

Currently, Mars is a hyperarid, hypothermal desert and the largest reservoir of surficial water ice on Mars resides at the poles. It is known, however, that variations in spin-axis/orbital parameters (obliquity, eccentricity, and precession; Laskar et al., 2004) can cause mobilization of water ice, transport in the atmosphere, and redeposition at lower latitudes. Evidence has been presented that a recent, meters-thick, ice-rich mantle was emplaced from the poles down to about 30° N and S latitude in the last several million years during an "ice age," and that it has been undergoing modification in the 30°-50° degree latitude region in the last few hundred thousand years, as Mars' obliquity amplitude decreased and ice returned to the poles (Head et al., 2003). During earlier periods of higher obliquity (mean obliquity of ~35°) in the Late Amazonian, ice was deposited in the mid-latitudes, and formed widespread valley and plateau glacial landsystems

(Madeleine et al., 2009; Head et al., 2010). Somewhat earlier in the Amazonian, during periods when the mean obliquity was thought to have been $\sim 45^\circ$, vast tropical mountain glaciers formed on the flanks of the major Tharsis volcanoes (Head and Marchant, 2003; Shean et al., 2005, 2007; Milkovich et al., 2006; Kadish et al., 2008a).

Unknown, however, has been the exact pathway and residence time of volatiles during transitions from one regime to another. Geological evidence, for example, has suggested that the higher amplitude obliquity of the past few million years caused ice stability conditions to migrate equatorward, and resulted in the deposition of a dust-ice mixture as a broad circum-polar high latitude mantle during periods of high obliquity (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2003). In contrast, atmospheric general circulation models suggest that during periods of higher obliquity (mean obliquity $\sim 25^\circ$ and high amplitude variation) ice migrates directly to equatorial regions and then works its way back to the mid to high latitudes to be deposited in a more stable environment (Levrard et al., 2004). In a similar manner, mid-latitude glaciation is best explained in Mars general circulation models if mean obliquity is $\sim 35^\circ$ and the source of ice is at the equator, not at the poles (Madeleine et al., 2009). However, direct evidence of the presence of large quantities of ice that could serve as equatorial sources in the recent geologic past has not yet been documented. Finally, at mean obliquity of $\sim 45^\circ$, ice is predicted to be deposited directly in the equatorial regions and to

remain there as long as these conditions prevail (Forget et al., 2006), accumulating sufficient ice to produce the observed tropical mountain glaciers (e.g., Head and Marchant, 2003).

How can the differences between the equatorial ice predicted to occur by the models, and the lack of geological observations for the presence of such deposits, be reconciled? One difficulty recognized by both geologists and climate modelers is that ice mantles and glacial deposits are destined to be cold-based in the hypothermal, hyperarid Mars environment (Marchant and Head, 2007). Cold-based ice deposited at high polar latitudes on Earth and over much of Mars does not erode its substrate in any substantial manner (Marchant and Head, 2007) and thus contains little to no debris unless such debris is deposited from the atmosphere or falls on the ice surface from adjacent high topography exposing cliffs of rock and debris. Thus, for typical cold-based deposits without debris, when the climate changes from mean accumulation to mean ice loss, the ice deposits are destined to sublimate, returning to the atmosphere and leaving no geological record. On the other hand, if the cold-based ice contains debris from adjacent cliffs (Marchant et al., 2002; Head et al., 2010), tephra from adjacent volcanoes (Shean et al., 2007; Wilson and Head, 2009), atmospheric dust co-deposited with ice (Head et al., 2003), or is armored by processes related to impact cratering (Kadish et al., 2008b; Schaefer et al., 2011), then evidence of the presence of preexisting cold-based ice can be preserved.

One well-documented example of this type of preservation of evidence for cold-based ice occurs in the form of pedestal craters, a unique crater form that was first identified in Mariner 9 data (McCauley, 1973). Initially pedestal craters were thought to result from eolian deflation of fine-grained intercrater material surrounding armored pedestals (McCauley, 1973; Arvidson et al., 1976). More recent studies have favored impact craters that formed in ice-rich substrates in conjunction with armoring that inhibits sublimation for higher latitude pedestal craters (e.g., Wrobel et al., 2006; Kadish et al., 2008b, 2009, 2010a; Nunes et al., 2011). The ice rich substrate in which an impact crater formed has subsequently eroded in the region, leaving a crater perched on a plateau or mesa-like pedestal of material that terminates at an outward facing scarp located up to several crater radii from the rim crest (Barlow et al., 2000). Formation of pedestal craters requires an armoring mechanism induced by, or related to, the impact event that improves the preservation potential of the impact substrate in the immediate vicinity of the crater relative to unperturbed areas. Both pedestal substrates (material composition) and armoring mechanisms have been the subject of extensive investigation (e.g., Head and Roth, 1976; Mouginis-Mark, 1987; Schultz and Mustard, 2004; Barlow, 2006; Osinski, 2006; Boyce et al., 2008). In summary, following crater formation in the ice-rich substrate, when climate conditions changed sufficiently to cause regional ice loss, the intervening ice deposits between pedestal craters disappeared, but the

armorings protected the ice in the substrate around the crater, leaving the pedestal crater as evidence of the presence and thickness of the ice deposit.

During an analysis of the nature of very young impact craters (Schon and Head, 2011) in relation to the latitude-dependent ice-rich mantle (Head et al., 2003), we discovered a population of very small (decameter-sized) pedestal craters in the tropics of Mars, superposed on ejecta from a 5.3-km diameter crater that formed about 12-13 million years ago. While km-scale mid-latitude pedestal craters are likely to have formed over Amazonian climate epochs of ~100 Myr duration (Kadish et al., 2009), because of their substantially smaller size, the newly observed pedestal craters examined here are sensitive to meters-scale substrates (now substantially removed) that are of keen interest because of their potential relation to late Amazonian climate conditions and volatile transport pathways. Here we present evidence that these small pedestal craters formed when a meters-thick layer of ice was present in the tropics of Mars in the last few million years. These features provide the first observational evidence that an ice reservoir existed in the tropics in the very recent geological history of Mars.

Observations

In the Daedalia Planum region of Mars (23°S, 230°E) on the Tharsis rise (Fig. 1), decameter-scale pedestal craters are observed superposed on the ejecta deposit of a recent 5.3 km-diameter crater (Fig. 2). These features (Fig.

3) are morphologically similar to larger pedestal craters, but occur at a low-latitude location that is significantly equatorward of typical higher-latitude pedestal craters (poleward of $\sim 40^\circ\text{S}$). Commensurate with their smaller diameters, these pedestal craters have much thinner pedestals.

Decameter-scale pedestal craters: Unambiguous examples of decameter-scale pedestal craters (Fig. 2, Fig. 3) are observed ranging in diameter from 7.9 meters to 29.5 meters with pedestals extending from 1.9 to 3.2 crater radii beyond the rim crest (Table 1). Shadow measurements were used to estimate crater depths and pedestal thicknesses (Table 1). Given a pedestal thickness of one to several meters, the crater depths (1.6 – 4.6 m) suggest that these craters have excavated through the pedestal substrate and into the underlying material. Depth-to-diameter ratios (Table 1) are consistent with the typical value of ~ 0.20 for simple craters on Mars (Strom et al., 1992; Garvin et al., 2003). Pedestal to crater radius (P/C) ratios (Table 1) are comparable to reported mean values for mid-latitude km-scale pedestal craters: ~ 3.3 for the northern mid-latitudes and ~ 2.5 for the southern mid-latitudes (Kadish et al., 2009). Values for pedestal circularity (Table 1) exceed the mean value of 1.10 reported by Kadish et al. (2009), but are less than the mean pedestal circularity of 1.64 observed in the Medusae Fossae Formation (Kadish et al., 2009). Pedestal volume to crater volume estimates (Table 1) were calculated by assuming homogenous thicknesses for the pedestals and

parabolic shapes for the craters. These estimates indicate that as expected pedestal volumes substantially exceed crater volumes (Table 1).

The pedestals range from symmetric about the crater (e.g., Fig. 3A) to more asymmetric (e.g., Fig. 3E), but the morphology of the pedestal margins does not suggest strong control by a prevailing wind regime (Fig. 3). Pedestal boundary scarps are commonly sharp and well defined (Fig. 3A, 3C, 3D, 3G), to locally more rounded (Fig. 3B, 3E). Pedestal surfaces appear smooth (e.g., Fig. 3F, 3G) or modestly stippled (e.g., Fig. 3A, 3C). Figure 3B, 3C, and 3D contain small craters or pits on the pedestals, which may suggest that these impacts were clusters (Ivanov et al., 2008). Elevated rim crests are common (e.g., Fig. 3A, 3C, 3D, 3H). Crater interiors appear partially filled (Fig. 3). For example, the three largest examples (Fig. 3A, 3C, and 3D) contain materials with a dune or pit-like texture, while the pedestal crater in Fig. 3I contains texturally smooth material.

Chronological Constraints: The regional terrain underlying the pedestal craters is comprised of Amazonian lava flows emerging from the Arsia Mons region (Scott and Carr, 1978) of the Tharsis rise (Fig. 1). Locally, the pedestal craters sit on top of the northern flank of an ejecta deposit of a young 5.3 km-diameter crater (Fig. 2). Using the CraterTools extension for ArcGIS (Kneissl et al., 2011) we counted impact craters superposed on the ejecta deposit using sub-meter resolution HiRISE data (McEwen et al., 2007). A 100-m grid was implemented using Hawth's tools (Beyer, 2004) to facilitate

systematic crater counting. The crater count methodology has been criticized based on interpretations that secondary craters, which result from the fallback of primary crater ejecta, are overwhelming at small sizes (McEwen et al., 2005). However, the crater counting system employs all scattered craters, including secondary craters, and recent tests have supported the system (Hartmann, 2005; Hartmann et al., 2010). Direct observations of the impact rate (Malin et al., 2006; Daubar et al., 2010) are also consistent with predictions from the isochrons (Hartmann, 2007; Kreslavsky, 2007).

On the basis of our count data, the crater size-frequency distribution observed on this deposit (Fig. 4A) suggests a best-fit age of ~ 12.5 Ma calculated using craters with diameters larger than 8 m. An apparent deficit of craters less than approximately 10 meters in diameter occurs compared to the isochrons of Hartmann (2005). This may be attributable in part to the difficulty identifying small impact craters in relatively rough and pitted regions of the deposit. In addition small craters could also have been removed or filled by the emplacement and removal of the layer responsible for the ~ 8 -30 m diameter pedestal craters. Degraded and partially filled craters are also observed, which is consistent with the removal of a surface layer. On the basis of the ~ 12.5 Ma age of the ejecta, and the superposition of the pedestal craters, deposition of the pedestal substrate and formation of the pedestal craters by preferential removal of the intercrater material must post-date the ~ 12.5 Ma ejecta deposit.

Could the pedestal craters be substantially younger than this age? The extremely small surface areas of the pedestals considered here prevent their analysis using conventional crater retention age techniques to count superposed craters and derive ages as has been done for larger pedestal craters (e.g., Kadish et al., 2010b). However, the crater-size frequency distribution represented by the population of pedestal craters themselves (Fig. 4B) provides information on the timescale of their formation, but does not constrain the exact temporal occurrence of this period. These data suggest that the pedestal craters formed during an interval or intervals that post-dated the ~12.5 Ma ejecta unit and collectively lasted at least 600 kyr, the "age" represented by the portion of superposed impact craters that are pedestal craters. Of course, the pedestal craters could have formed over a singular ~600-kyr period, or over a set of shorter, recurring intervals when episodic ice deposits were present.

Discussion

The craters described here (Fig. 3) exhibit morphological attributes, such as pedestal extent and terminal scarp, commonly used to define pedestal craters (Barlow et al., 2000; Wrobel et al., 2006; Kadish et al., 2009). Higher latitude pedestal craters in the southern hemisphere are characterized by pedestals that are tens of meters (mean = 20.4 m, median = 35.0 m) thick in which the crater is perched (the crater does not typically penetrate beyond

the pedestal substrate) (Kadish et al., 2010a). In contrast, these decameter-scale pedestal craters penetrate through the pedestal substrate and are similar in this fashion to excess ejecta craters (Black and Stewart, 2008). Given their uniquely small size and low latitude (23°S), what is the composition and origin of the substrate for these pedestal craters?

Recent advances in multiple methods of remote sensing, climate modeling, and observation show that subsurface ice, glacial accumulations of ice, and pervasive ice sheets have all been common in the Amazonian. Superposed and draped pedestal craters attributed to cyclically-emplaced ice-rich deposits (Kadish et al., 2010a, 2010b) and observations of marginal pits (Kadish et al., 2008b) are consistent with the preservation of volatile material under an indurated pedestal surface that has resisted erosion and retarded vapor diffusion relative to intercrater material. Climate modeling studies and geological analyses suggest that significant mobilization and redistribution of ice is controlled by orbital and spin-axis parameters. For example, in 45° obliquity simulations, Forget et al. (2006) has shown that ice precipitates at equatorial and mid-latitude locations where extensive geological evidence of glaciation is observed (e.g., Head et al., 2005).

Gamma-ray and neutron spectroscopy have revealed the latitude-dependence of very near surface ice (Boynton et al., 2002). These data support interpretations of surface roughness trends that a meters to tens of meters thick mantle has smoothed high-latitude topography (Kreslavsky and

Head, 2000). The morphology of this deposit has been interpreted to be the result of deposition of dusty snow during a recent ice age (Mustard et al., 2001; Head et al., 2003) and stratigraphic layers have been observed along mid-latitude erosive margins (Schon et al., 2009). At higher latitudes, pervasive polygons suggest a very ice-rich composition (Levy et al., 2010), which has been observed directly following meters-scale impacts that expose nearly pure ice beneath a surface sublimation lag (Byrne et al., 2009). In addition to these lines of evidence for extensive past glaciations and present subsurface ice in the mid-latitudes, recent near-infrared observations of the seasonal stability of carbon dioxide ice deposits on the surface imply that near-surface ice is modifying thermal inertia (Vincendon et al., 2010). Importantly, these observations show that shallow buried water ice is currently present on appropriate slopes at latitudes as low as 25° in the southern hemisphere. Such shallow reservoirs of mid-latitude ice, likely deposited during ice ages (Head et al., 2003) of the previous few to tens of millions of years, suggest that a volatile-rich substrate is a plausible pedestal crater substrate at this locality (Fig. 1; Fig. 2).

Preservation of shallow ice in mantling deposits requires the development of a protective sublimation lag to retard vapor exchange with the atmosphere. Sublimation pits observed in areas of degraded mantle are evidence of this phenomenon (Milliken et al., 2003). Terrestrially, the development of a till cover has preserved glacial ice for more than 8 Myr in

Beacon Valley, Antarctica (Sugden et al., 1995; Kowalewski et al., 2011). On Mars, the dust cycle is expected to be more vigorous during periods of high obliquity (Haberle et al., 2003) and these dusty materials are likely to compose the very fine-grained component of mantling materials.

Furthermore, similar types of deposits are predicted on the sides of the Tharsis rise as relatively damp polar air migrates up the side of the Tharsis rise during periods of high obliquity, adiabatically cooling and depositing ice and dust mixtures (Forget et al., 2006). Additionally, this locality (Fig. 1) is located ~900 km from the Arsia Mons caldera, a potential source of tephra (Wilson and Head, 2009). Even without an eruption, the large volcanic edifice represents a proximal source of dusty material for regional redistribution and incorporation into mantling deposits. Radar sounding of higher latitude pedestals (Nunes et al., 2011) has revealed a dielectric permittivity consistent with a mixture of ice and silicate materials (or a highly porous ice-free material such as volcanic ash).

Additional observations of craters preserving evidence of ice-rich substrates are provided by Black and Stewart's (2008) evaluation of km-scale excess ejecta craters, which they define as craters for which the volume of ejecta deposit material above the background surface substantially exceeds (>2.5x) the cavity volume below the background surface. Four recently reported excess ejecta craters have volume of ejecta to volume of crater cavity ratios ranging from 4.4 to 28.5 (Kadish and Head, 2011). These craters lack

the characteristic boundary scarps of pedestal craters, but typically have distinct ejecta deposit extents. Because all excess ejecta craters have fresh morphologies, Black and Stewart (2008) discount cavity infilling and post-impact mantling of the crater and ejecta as explanations for their formation. In their proposed model, the pre-impact surface has one or more ice-rich layers. The excess ejecta crater penetrates through the ice-rich material and excavates underlying substrate, which is incorporated into the ejecta deposit. When ice-rich materials sublime and are removed in the region, the crater's ejecta deposit inhibits removal of the icy material it covers (Black and Stewart, 2008). Both pedestal craters and excess ejecta craters result from the presence of an ice-rich surface unit at the time of impact and the partial preservation of the ice-rich deposit by armoring of the surface or superposition of ejecta (Kadish and Head, 2011).

Interpretations and Conclusions

The decameter-scale craters (Fig. 3) analyzed here are morphologically similar to much larger pedestal craters documented at higher mid-latitudes. Pedestal heights are consistent with a surface layer a few meters thick. Crater depths indicate penetration through this surface layer and excavation of underlying material (Table 1). Pedestal volumes substantially exceed crater volumes (Table 1). In our preferred interpretation, this surface layer (now largely removed) resulted from climate variations that drove

accumulation of meters-thick ice deposits. Evidence for the cyclical deposition of such icy layers during the late Amazonian is well documented (Kreslavsky and Head, 2000; Mustard et al., 2001; Head et al., 2005; Forget et al., 2006; Schon et al., 2009) and has been interpreted as a series of ice ages (Head et al., 2003; Schorghofer, 2007). Climate modeling studies (e.g., Mischna et al. (2003), Levrard et al. (2004), Forget et al. (2006), and Madeleine et al. (2009)) have suggested that equatorial deposition preceded mid-latitude mantling events in the late Amazonian and our observations provide the first geological support for this scenario of recent equatorial ice associated with obliquity variations.

We propose that ejecta of the decameter-scale craters derived from underlying material provided the armoring mechanism to preserve a portion of the mantling layer. In our model, deposition of this icy and dusty mantling layer occurred on the ejecta deposit that we estimate as ~12.5 Myr old (Fig. 4A). The impact craters that occurred then penetrated through the icy surface layer. These craters excavated underlying material onto the surface layer. The surface layer endured for at least 600,000 years in one or more episodes (Fig. 4B) to accumulate the morphologically distinct craters that are observed. Ejecta of the craters protected the underlying surface layer from sublimation, dissection, and eolian erosion, which have stripped away the surface layer in the intercrater areas. In this fashion our model is similar to the model of excess ejecta crater formation proposed by Black and Stewart

(2008). An atmospheric blast effect as described by Wrobel et al. (2006) or another non-ballistic ejecta armoring mechanism (e.g., Schultz and Mustard (2004)), which appears necessary to explain larger fully-perched pedestal craters (e.g., Kadish et al., 2009), is not necessary to explain the more modest pedestal extents observed here (Table 1) due to the possible armoring effect of sub-ice excavated ejecta (e.g., Kadish and Head, 2011).

When was the icy layer deposited and when did the pedestal craters form? We cannot know for certain, because there could be successive periods of deposition and removal during which normal craters were formed. Our morphological observations (Fig. 2; Fig. 3) and chronological constraints (Fig. 4), however, in conjunction with the obliquity history calculated by Laskar et al. (2004), provide a framework for considering hypotheses. One possibility is that the pedestal craters could have formed during the most recent $\sim 0.4 - 2.1$ Ma ice age (Head et al., 2003). However, we suggest that formation is more likely to have preceded this most recent ice age and occurred during a period of higher mean obliquity prior to 5 Ma, during which obliquity excursions often approached 45° (Fig. 5). These older more extreme obliquity conditions are consistent with global climate modeling results of equatorial ice deposits at 45° obliquity (e.g., Forget et al., 2006). Additionally, the longer succeeding time interval is consistent with the destruction and removal of the intercrater material by sublimation and eolian erosion during subsequent history (<5 Ma).

Global climate modeling studies reveal that high obliquity conditions ($\sim 45^\circ$) are required to construct the large tropical mountain glaciers associated with the Tharsis Montes (Forget et al., 2006). The geological records of these significant tropical mountain glaciers (Fig. 1), for which climate models indicate sufficient ice deposition at high obliquity (Forget et al., 2006), are substantially older (~ 65 Ma and older, Shean et al., 2007) than more recent obliquity peaks (Fig. 5). Reconnaissance of the Arsia Mons fan-shaped deposit reveals small-scale pedestal craters in Mars Reconnaissance Orbiter Context Camera (Malin et al., 2007) images (Fig. 6) similar to those documented at Daedalia Planum (Fig. 3). These craters (Fig. 6) are superposed on the ridged facies of the fan-shaped deposit (interpreted as drop moraines; Head and Marchant (2003); Shean et al. (2007)). We interpret these craters as geological evidence that less voluminous quantities of ice have been present in the Tharsis region substantially more recently than indicated by the underlying Amazonian deposits of the large tropical mountain glaciers (Head and Marchant, 2003; Shean et al., 2005, 2007; Milkovich et al., 2006; Kadish et al., 2008a). For these larger (~ 100 -m diameter) pedestal craters, higher P/C ratios and values of pedestal circularity are observed (Table 2). The P/C ratios are more comparable to the mean P/C ratio of 5.6 observed by Kadish et al. (2008b, 2009) for pedestal craters with marginal pits in Utopia Planitia and Malea Planum that are interpreted as sublimation features.

Together, these newly identified small pedestal craters (Fig. 3; Fig. 6) imply a greater equatorial extent of surficial ice deposits in the equatorial and mid-latitudes during latest Amazonian climate variations (Fig. 5) than previously known from geological observations. While high-latitude mantling with polygonal patterning is quite young (Kostama et al., 2006; Levy et al., 2010), the precise chronology and latitudinal extent of previous glacial events has been more uncertain (e.g., Head et al., 2003; Schon et al., 2009). We suggest that the substrate for these small pedestal craters was an ice-rich layer that formed, perhaps intermittently, during the period of higher obliquity that prevailed prior to 5 Myr ago (Fig. 5). These pedestal craters imply that surficial ice-rich mantling deposits were present in the tropics during the previous ~12.5 Ma and have since eroded back to the armored pedestal craters. These observations expand the latitude range over which such deposits are known to have occurred and constrain their deposition to the latest Amazonian (Fig. 5). These deposits represent a plausible source for recent detections of shallow ice at low mid-latitudes (e.g., Vincendon et al., 2010). Finally, these results support the predictions of general circulation models that ice migrated to the equatorial regions during recent periods of enhanced obliquity and from there served as a source for mid-latitude ice-rich mantling deposits.

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Table 1: Figure 3 pedestal crater measurements. Crater depths and pedestal thicknesses are estimated from shadow measurements.

Panel (Fig. 2)	Crater Diameter (m)	Crater Depth (m)	d/D	Max. Pedestal Extent (m)	Pedestal Thickness (m)	P/C ratio ¹	Pedestal Circularity ²	V _{pedestal} / V _{cavity}
A	29.5	4.6	0.15	43.0	1.1	2.92	1.45	4.4
B	8.1	1.6	0.19	9.9	1.1	2.44	1.37	7.5
C	28.1	3.6	0.13	42.2	2.0	3.00	1.43	7.5
D	20.2	3.6	0.18	22.5	1.5	2.23	1.36	4.1
E	10.3	1.9	0.18	10.3	1.3	2.00	1.24	4.2
F	8.5	1.6	0.19	11.6	1.0	2.73	1.26	6.3
G	12.9	1.8	0.14	12.5	1.3	1.94	1.19	4.5
H	10.8	2.5	0.24	10.5	1.1	1.94	1.20	3.2
I	13.7	1.8	0.13	18.6	1.2	2.72	1.19	11.4
J	7.9	1.9	0.24	12.7	0.8	3.22	1.60	7.7

¹The pedestal to crater radius ratio, P/C Ratio = (farthest extent of pedestal) / (crater radius).

²Pedestal circularity = (pedestal perimeter) / (4 π (pedestal area))^{1/2}

Table 2: Pedestal craters observed on the Arsia Mons fan-shaped deposit (Figure 6).

Crater (Fig. 6)	Location	Crater Diameter (m)	Max. Pedestal Extent (m)	P/C ratio ¹	Pedestal Circularity ²
A	2.82°S, 231.42°E	118	450	7.63	1.73
B	2.80°S, 230.72°E	93	231	4.97	1.71
C	4.62°S, 230.79°E	125	263	4.21	1.41

¹The pedestal to crater radius ratio, P/C Ratio = (farthest extent of pedestal) / (crater radius).

²Pedestal circularity = (pedestal perimeter) / (4 π (pedestal area))^{1/2}

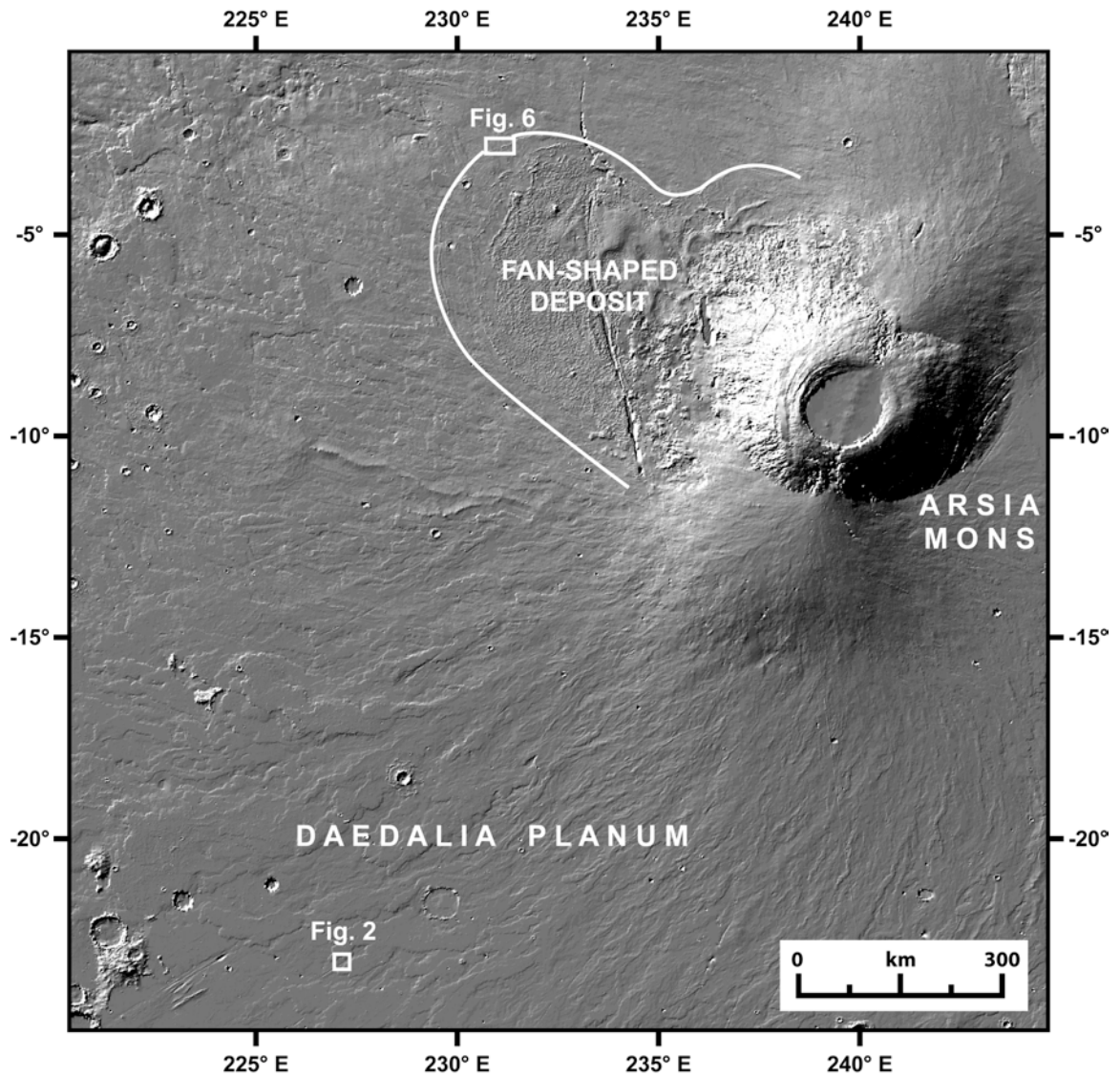


Figure 1: The Arsia Mons and Daedalia Planum region. The northwest flank of Arsia Mons hosts a ~166,000-km² fan shaped deposit interpreted as evidence of cold-based mountain glaciers. Daedalia Planum is an Amazonian volcanic plain on the southwest side of the Tharsis rise; lavas are derived from Arsia Mons.

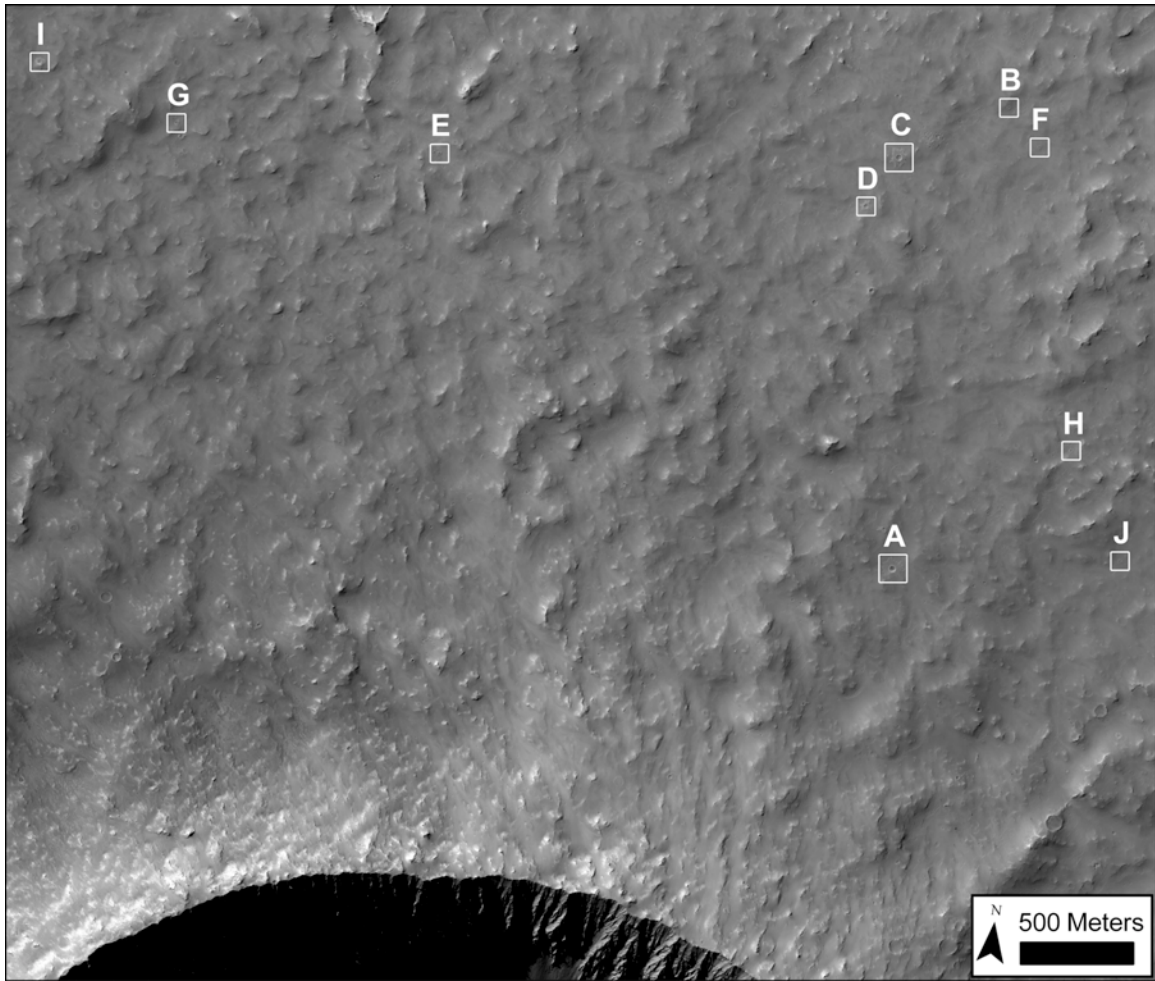


Figure 2: The ejecta-deposit of a recent 5.3 km-diameter impact crater (23° S, 230° E) (rim crest visible at bottom) underlies decameter-scale pedestal craters observed in the Daedalia Planum region. The locations of pedestal craters shown in Figure 3 are labeled here with corresponding letters. Portion of HiRISE: PSP_007735_1570.

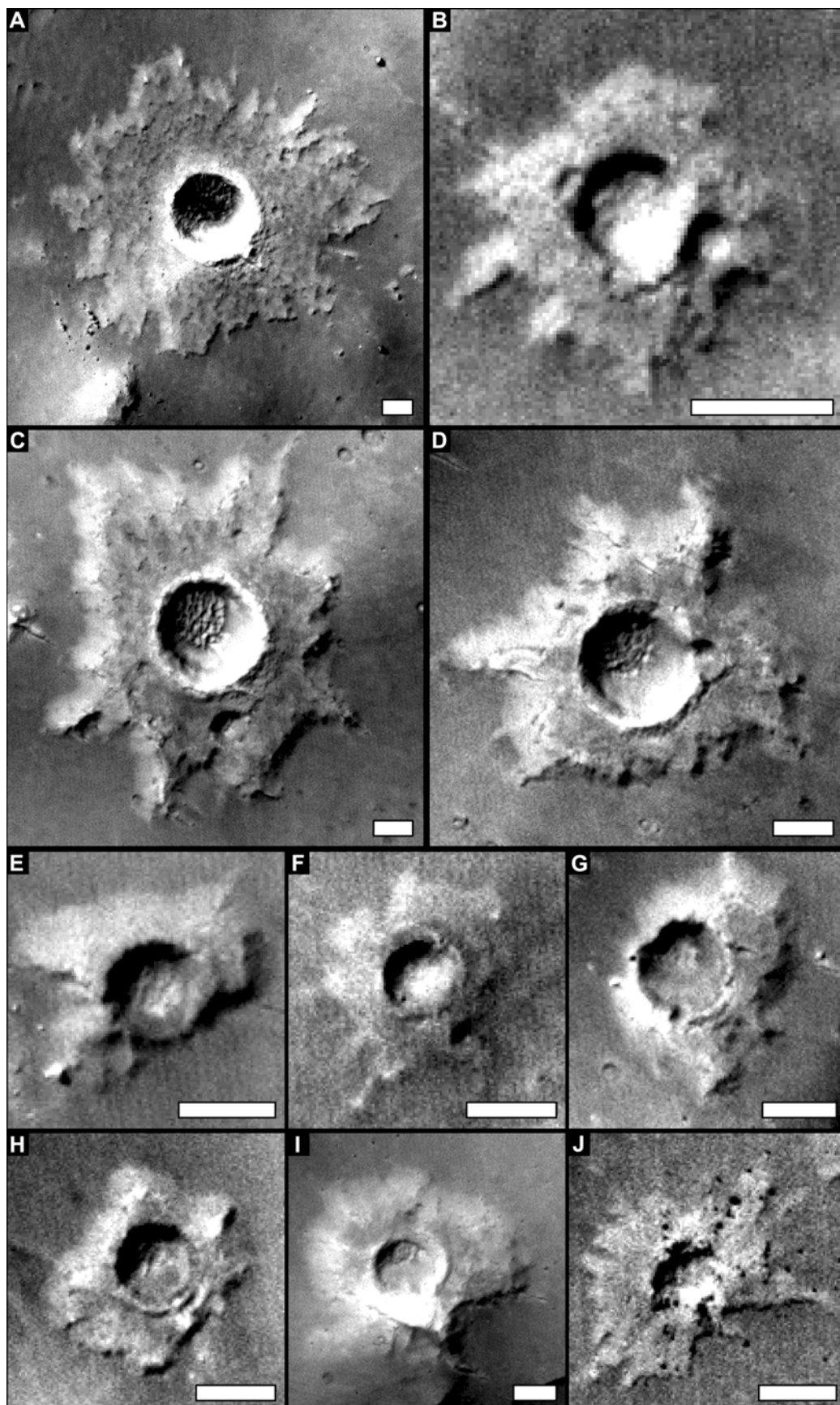


Figure 3: Decameter-scale pedestal craters observed in the Daedalia Planum region (23° S, 230° E). The scale bar in each panel is 10 meters and north is always oriented up. Crater measurements are reported in Table 1. Portion of HiRISE: PSP_007735_1570.

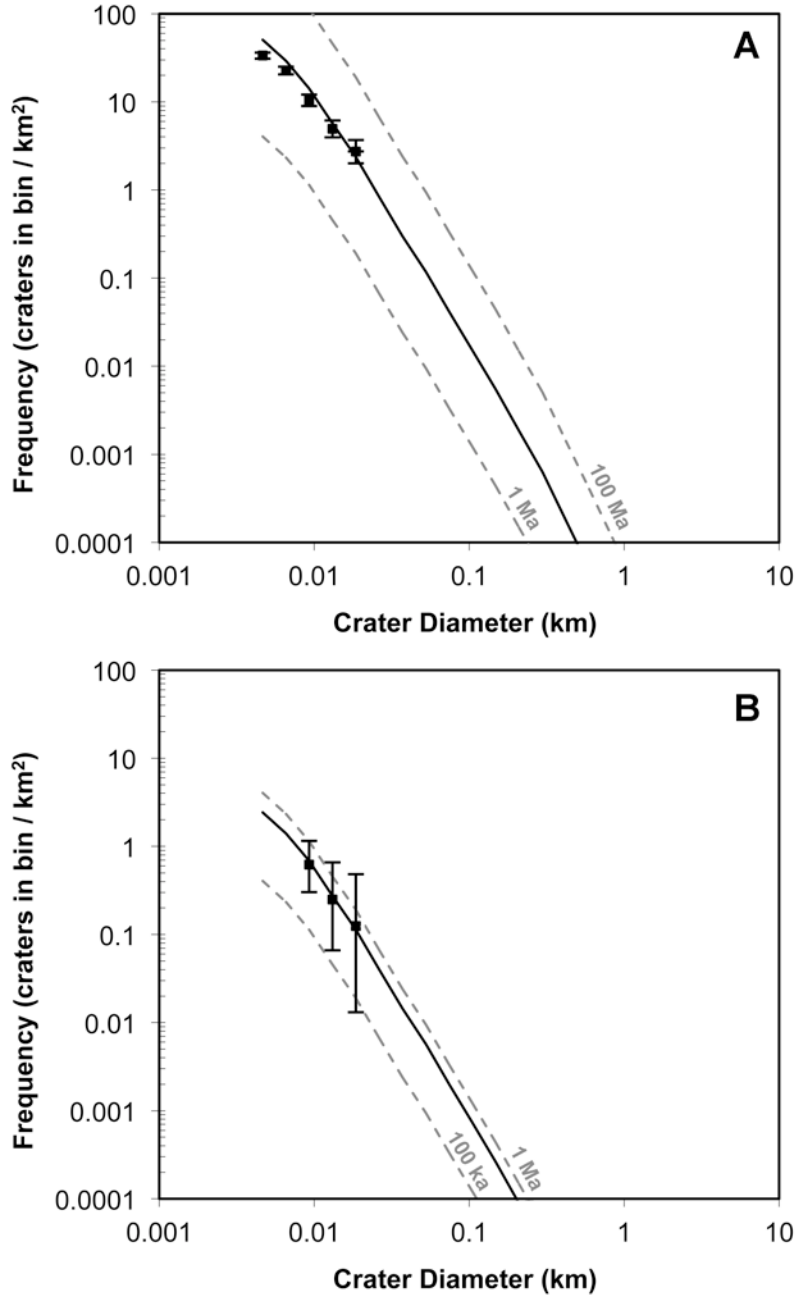


Figure 4: (A) A count of all craters on the ejecta deposit underlying the pedestal craters suggests an emplacement date of approximately 12.5 Ma. This age is a best fit derived using a minimum crater diameter of 8 meters and isochrons of Hartmann (2005). The count area was 7.66 km² of HiRISE: PSP_007735_1570 and 1631 craters were counted. A deficit of craters below

~0.01 km in diameter is partially attributed to difficulty distinguishing craters in this size range in areas of pitting and to the preferential removal of craters smaller than those in Figure 3 with the removal of the surface layer. Degraded and partially filled craters are also observed, which we interpret as consistent with the removal of the surface layer. Emplacement of the pedestal crater substrate and formation of the pedestal craters must post-date this age. (B) This crater size-frequency distribution of only the pedestal crater population suggests a formation timescale of approximately 600 kyr, but does not constrain pedestal crater formation to a single episode of this duration. These data are a minimum for the age of the pedestal craters. The age cannot be constrained using conventional crater retention age techniques due to their very small surface areas (e.g., Kadish and Head, 2010b).

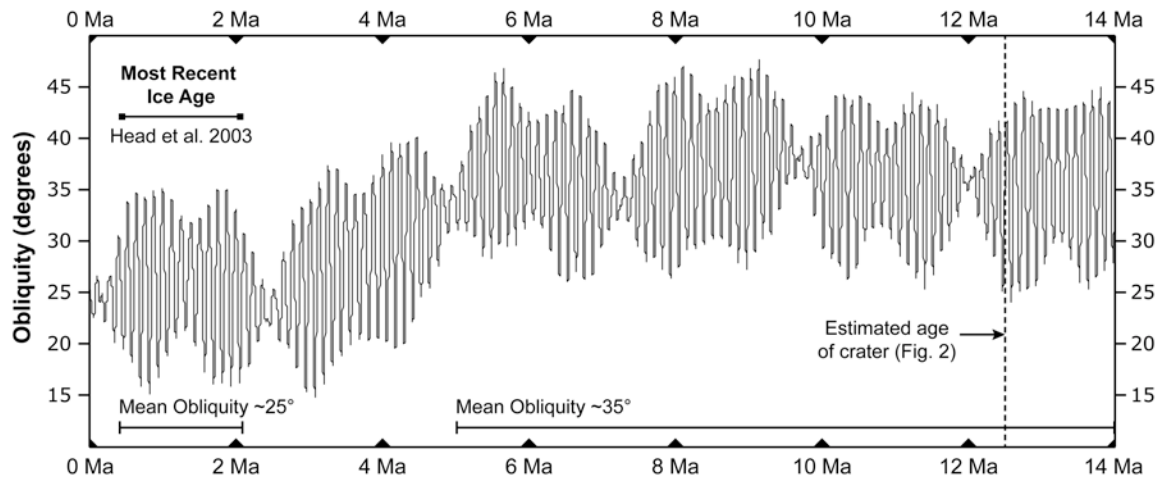


Figure 5: Mars recent obliquity (Laskar et al., 2004). Enhanced obliquity variations, 0.4 Ma to 2.1 Ma, characterize what has been interpreted as Mars’ most recent “ice age” (Head et al., 2003). In our interpretation, the pedestal crater substrate was deposited during the period of higher obliquity prior to 5 Ma.

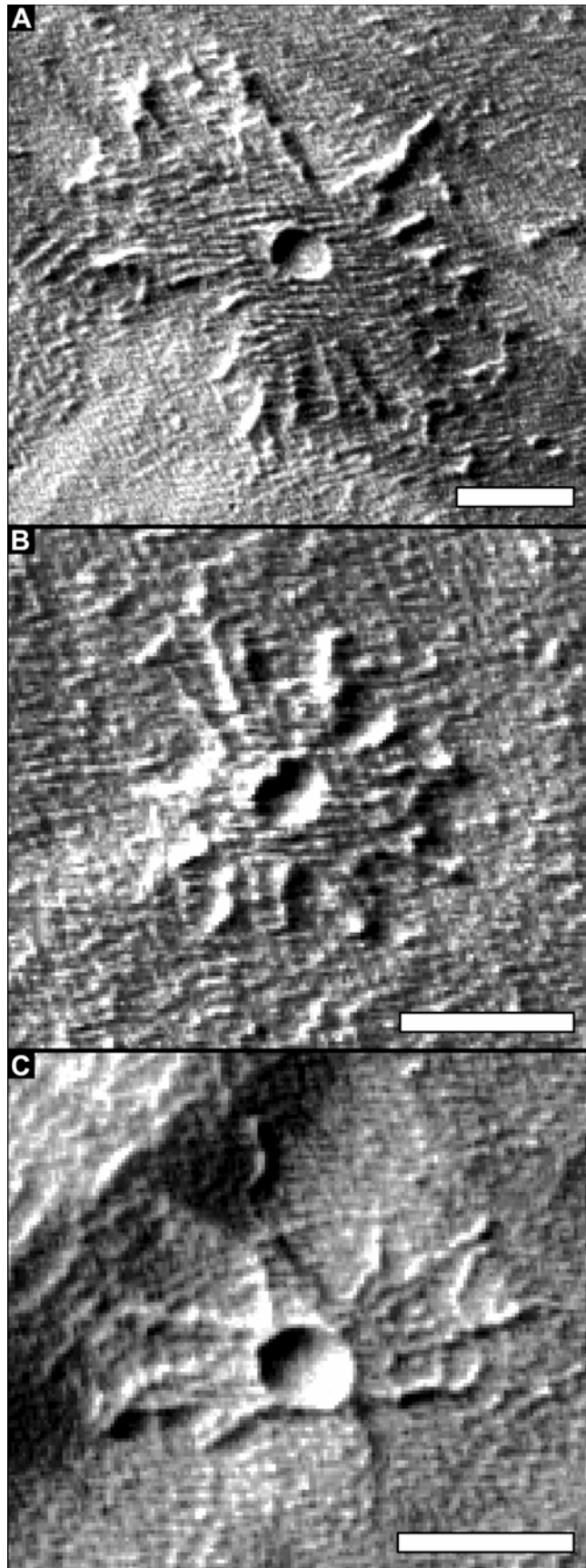


Figure 6: Pedestal craters on the Arsia Mons fan-shaped deposit. The northwest flank of Arsia Mons hosts a ~166,000-km² fan shaped deposit that is bounded by parallel ridges interpreted as drop moraines from cold-based glaciers (Head and Marchant, 2003; Shean et al., 2005; Marchant and Head, 2007). The location of this large glacial deposit is consistent with GCM predictions of ice deposition at high obliquity (Forget et al., 2006) and ice sheet models of glacier evolution (Fastook et al., 2008). Reconnaissance in this region revealed small pedestal craters that are morphologically similar, superposed on the glacial deposit (A and B) and the moraines (C). The occurrence of these pedestal craters at this location is consistent with climate model predictions of both earlier, high-obliquity formation of tropical mountain glaciers, and much later, transient, much more modest equatorial surface ice deposition during more recent obliquity excursions (Fig. 5). The scale bar in each panel is 250 meters and north is always oriented up. Crater measurements are reported in Table 2. Portions of CTX: P21_009251_1771_XI_02S128W (A and B); P19_008605_1772_XI_02S129W (C).

CHAPTER 6

RECENT HIGH-LATITUDE RESURFACING BY A CLIMATE-RELATED LATITUDE-DEPENDENT MANTLE: CONSTRAINING AGE OF EMPLACEMENT FROM COUNTS OF SMALL CRATERS.

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Abstract

A surficial mantling deposit composed of ice and dust blankets Mars at high and mid-latitudes. The emplacement and evolution of this deposit is thought to be driven by astronomical forcings akin to, but more extreme than, those responsible for Earth's ice ages. In order to test predictions about the age of this deposit, more than 60,000 superposed craters were counted on near-rim terrain of 16 young craters using sub-meter resolution images. A chronology for the deposition of the latitude-dependent mantle is revealed by these data and shows that: (1) the overall age and age trend of mantling deposits is consistent with first-order control by obliquity variations; (2) mantling processes are substantially younger than some equatorial rayed craters which have crater retention ages of ~20-30 million years; (3) the mantle is younger by a factor of several at polar latitudes compared to its

furthest equatorial extent ($\sim 30^\circ$). Additionally, these crater counts confirm the suitability of small craters in geological analyses of youthful planetary surfaces and provide new data on the density of meters-scale craters superposed on deposits of geologically-recent craters.

Introduction

The Mars Orbiter Laser Altimeter (MOLA) revealed progressive topographic smoothing in trends of surface roughness from mid- ($\sim 30^\circ$) to high- ($>60^\circ$) latitudes, suggesting the effect of a meters to tens of meters thick blanketing deposit pervasive at high latitudes and degraded at mid-latitudes (Kreslavsky and Head, 2000). Morphological observations of both intact (smooth) and dissected (pitted and/or knobby) terrain are interpreted as evidence of a surficial ice-rich deposit that mantles pre-existing terrain and was recently subject to degradation (Mustard et al., 2001). These and related observations led to a model of mantle formation from dusty snow deposited during an obliquity controlled ice age (Head et al., 2003), a hypothesis consistent with stratigraphic layers observed along mid-latitude erosive margins of the mantling deposit (Schon et al., 2009a).

Additional remote sensing measurements and geomorphic evidence of the atmospherically-deposited ice-rich mantle have accumulated over the past few years (Head et al., 2011). These include gamma-ray and neutron spectroscopy, polygonally patterned ground, observations by the Phoenix

lander, observations of fresh ice exposure by impact craters and subsequent sublimation, mid-latitude gullies, and visible/near-infrared observations of carbon dioxide frosts. Gamma-ray and neutron spectroscopy have revealed the latitude-dependence of very near surface ice (e.g., Boynton et al., 2002), including at volumes improbable via vapor diffusion alone. Consistent with the ice contents reported by Boynton et al. (2002), meter-scale polygons at these latitudes are interpreted to be thermal contraction crack polygons that require a very ice-rich substrate for their formation (e.g., Levy et al., 2008). In landing with retrorockets, the Phoenix spacecraft (68.23°N, 234.25°E) uncovered homogenous smooth ice (“Snow Queen”) under the lander (Smith et al., 2009).

Elsewhere on Mars, repeat imaging by the context camera, followed by targeted HiRISE imaging, has led to the identification of new impact craters that reveal a nearly pure ice substrate beneath a sublimation lag on the surface (Byrne et al., 2009). Sublimation of exposed ice has been observed and comparison of these observations to sublimation models supports a nearly pure water ice composition for the material (Dundas and Byrne, 2010). In the mid-latitudes, gullies (Malin and Edgett, 2000) are a prevalent landform on steep, especially pole-facing, slopes (e.g., Dickson et al., 2007). Morgan et al. (2010) has shown that mid-latitude gully morphology is sensitive to slope orientation and insolation conditions, thereby suggesting that top-down melting of surficial snow and ice deposits associated with the

latitude-dependent mantle is the meltwater source for gully formation. Such a surficial meltwater source is similar to previous interpretations of surficial ice melting (e.g., Costard et al., 2001; Christensen, 2003), and points to an intimate genetic relationship between ice-rich mantling deposits and mid-latitude gully development (Head et al., 2008; Dickson and Head, 2009; Schon and Head, 2011a). Finally, recent near-infrared observations of the seasonal stability of carbon dioxide ice deposits on the surface imply that near-surface ice is modifying thermal inertia in mid-latitude regions (Vincendon et al., 2010). Collectively, these observations provide significant evidence in support of widespread and relatively young ice-rich mantling deposits that blanket mid- and high- latitudes (Head et al., 2011), but they do not constrain the absolute age of the deposits, or episodes of emplacement. Models of vapor diffusion (e.g., Mellon et al., 2004; cf., Bandfield, 2007) can characterize the present-day stability of buried ice, but also cannot directly shed light on the age or origin of buried ice sheets.

How can the absolute age of this mantle be determined? Emplacement of the latitude-dependent mantle has been described as a martian ice age, driven by variations in orbital and spin-axis parameters, dominantly obliquity (Head et al., 2003). Large variations in Mars obliquity (tilt of the planetary axis of rotation relative to the orbital plane) have long been recognized as having a large influence on martian (paleo)climate (Ward, 1973; Sagan et al., 1973). These variations are chaotic (e.g., Touma and

Wisdom, 1993), but robust numerical solutions for the recent geologic past (20 Ma to present) have been developed by Laskar et al. (2004). A period of enhanced obliquity oscillations from 2.1 Ma to 400 Kyr has been described as Mars most recent ice age, during which the observed latitude-dependent mantling was emplaced (Head et al., 2003; Levrard et al., 2004). In this paradigm, the previous ice age waned approximately 2.8 Ma (Head et al., 2003). Similarly, based on the obliquity record (Laskar et al., 2004), Schorghofer (2007) has performed simulations that suggest “forty major ice ages over the past five million years.” These advances and retreats of ice stability are of varying extents (Schorghofer and Aharonson, 2005; Schorghofer, 2007) with unknown preservation of geologic evidence, though layered margins (Schon et al., 2009a) certainly support multiple episodes of mantle emplacement.

The youthfulness of high latitude mantling is known from the pervasiveness of un-modified polygonally patterned ground (Kreslavsky et al., 2011). Initial efforts to date the mantle by Kostama et al. (2006) using superposed craters suggested a very young crater-retention age and a latitudinal trend toward older mantle surfaces equatorward. These studies lead to a series of questions: What is the absolute age of the LDM? Does mid-latitude ($\sim 30^\circ$) mantling date from the most recent period of enhanced obliquity variations (~ 2.1 to 0.4 Ma)? What is the age of the LDM surface at high latitudes, and does it differ from that at lower latitudes? In this

investigation we use new sub-meter resolution data from HiRISE (McEwen et al., 2007) to investigate the age and chronology of the ice-rich latitude-dependent mantle at both high latitudes ($>60^\circ$) and the mid-latitude margin (30°). As part of our study, we date several equatorial rayed craters, which provide additional confirmation regarding the applicability of using small craters to date young surfaces and the validity of the crater chronometry system (e.g., Hartmann et al., 2010).

Approach

In order to assess quantitatively the youthfulness of mantle surfaces, small superposed craters are used to calculate crater retention ages. The technique of dating martian surfaces using the density of superposed craters is long established. Only recently have meter-scale resolution and better image datasets enabled analysis of small craters, leading to slight refinement of the isochrons (Hartmann, 2005) and inclusion of atmospheric filtering of small bolides (Popova et al., 2003). Criticism has been leveled at the crater count methodology based on interpretations of secondary cratering (McEwen et al., 2005); secondary craters result from the fallback of ejecta blocks launched by primary impacts. In the robust debate that followed, additional tests have been shown to support the crater count-based isochron system (Hartmann et al., 2008). For example, crater counts of rayed craters yield ages internally consistent with impact-size recurrence intervals anticipated

from the isochrons (Hartmann et al., 2010). In the present study, we date additional rayed craters and also find ages consistent with the isochron system.

Additionally, direct observations of the impact rate (Malin et al., 2006; Daubar et al., 2010) are consistent with predictions from the isochrons (Hartmann, 2007; Kreslavsky, 2007). As developed, the crater counting system employs all scattered craters, both primaries and distant secondaries, which are very difficult to distinguish geomorphically (Calef et al., 2009). Obvious crater rays and secondary crater chains are excluded in order to measure “the age-dependent general global buildup of both large and small primary craters and secondaries that accumulate simultaneously as a background, but minimize the effects of nearby primary craters” (Hartmann, 2005). Enhanced uncertainties may arise in the use of small craters (Hartmann, 2005; Hartmann, 2007), but the crater count methodology remains a robust and useful tool to assess very young martian surfaces (e.g., Werner et al., 2009).

All count areas in this study are located on near-rim deposits of morphologically fresh craters with diameters between ~ 1 and several km and for which HiRISE data was available (Table 1). The topography of these areas was reset by the crater ejecta (e.g., Melosh, 1989). Therefore, the surface is of a homogeneous age and the crater retention age represents either the age of the crater, or any post-crater resurfacing history. In this fashion, crater

counts can be performed that confidently date superposed mantle surfaces as well as un-mantled ejecta deposits, which can constrain the emplacement history of the mantle.

Methods

Young craters were identified at various latitudes based on their geomorphic characteristics using THEMIS (Christensen et al., 2004), CTX (Malin et al., 2007), and HiRISE data (McEwen et al., 2007). All of the craters examined in this study are simple impact craters, ranging in diameter from 0.8 to 7.3 km (Table 1). Pristine craters this size are bowl-shaped with a crisp raised rim crest and distinct ejecta deposits (e.g., Strom and Croft, 1992). Ejecta may vary in morphology (e.g., Barlow and Bradley, 1990), but typically the continuous ejecta deposit is well-defined. At the highest latitudes where polygonally patterned ground is pervasive, the comparative distinctness of rim crests was used to distinguish young craters. Not all of these young craters exhibit rays in nighttime thermal infrared (nTIR) data. Thirteen craters imaged by HiRISE and located in the southern hemisphere were selected for this study, and for comparison, three rayed craters (identified by Tornabene et al., 2006) in the northern hemisphere. Locations of these craters and summary data are presented in Fig. 1 and Table 1. Crater counts were conducted exclusively on sub-meter resolution HiRISE data (McEwen et al., 2007). For each count a 100 m grid was implemented using Hawth's tools

(Beyer, 2004) to facilitate an exhaustive search of near-rim ejecta deposit terrain for superposed craters. The CraterTools ArcMap extension (Kneissl et al., 2011) was used to consistently measure crater diameters without distortion. All crater data is presented in the incremental-plot style of Hartmann with isochrons from Hartmann (2005).

Crater Count Data:

Presentation of the crater count data is organized geographically. First, the three rayed craters in the equatorial region are considered (Fig. 2; Fig. 3; Fig. 4). These craters are located far from regions of latitude-dependent mantling (Fig. 1). Second, two high latitude craters are discussed (Fig. 5; Fig. 6), which have pervasive polygonally patterned ground draping their interiors, near-rim areas, and the surrounding terrain. Finally, a collection of craters from the mid-latitudes is considered (Figs. 7-16). Variations in latitude, age, and the presence or absence of superposed mantling deposits at these locations provide constraints on the history of latitude-dependent mantling.

Equatorial Rayed Craters: The distinctiveness of lunar crater rays arises from compositional and maturity differences with the background terrain (e.g., Hawke et al., 2004). Thermal inertia (TI) differences with surrounding terrain (e.g., low TI rays) are thought to be responsible for the thermophysical distinctiveness of martian rays (McEwen et al., 2005;

Tornabene et al., 2006; Preblich et al., 2007). Therefore, rays are most apparent in nighttime infrared data (Christensen et al., 2004). The distribution of identified rayed craters (Tornabene et al., 2006) suggests that the occurrence (or persistence) of rays is dependent on substrate.

Intermediate to high background thermal inertia and intermediate albedo appear to be important criteria regarding the distinguishability of rays (see global maps of Mellon et al., 2000 and Putzig et al., 2005). This is consistent with most of the Tornabene et al. (2006) detections occurring in equatorial volcanic terranes.

Thila (Fig. 2), Naryn (Fig. 3), and Dilly (Fig. 4) all preserve crater rays that are visible in thermal infrared data (Tornabene et al., 2006). These morphologically fresh craters are located in the equatorial regions near Elysium (Fig. 1). Detailed crater counts on near-rim deposits to date the formation of these craters (Fig. 2B, 3B, 4B) reveal ages that vary by more than a factor of ten. The crater retention ages of Thila, Naryn, and Dilly are 23.1, 2.1, and 34.4 Ma respectively. A study by Hartmann et al. (2010) estimated the age of Naryn as “a few Myr to 20 Myr.” These ages are consistent with very low erosion rates (Golombek and Bridges, 2000) and the general absence of very late Amazonian-aged glacial modification at these latitudes and longitudes (Kreslavsky and Head, 2006; Head and Marchant, 2009).

High-latitude Polygonalized Craters: In the higher latitude region where polygonally patterned ground is pervasive (e.g., Mangold, 2005; Levy et al., 2009), we examined two young craters and their near-rim deposits. One crater (Fig. 5) is located at 55.6°S, while the other (Fig. 6) is located substantially farther poleward at 77.9°S. At high latitudes, hectometer and kilometer-scale impact craters are not observed without a polygonal texture (Levy et al., 2009). The interior, walls, rim, and near-rim deposits of these craters are overprinted by a continuous polygonal pattern. These craters are considered youthful because of their comparatively distinct rim-crests and the presence of modest low-albedo regions surrounding them (Fig. 5A; Fig. 6A). Crater counts on the near-rim deposits of these craters are not indicative of the age of the crater; rather, the crater-size frequency distributions (Fig. 5B; Fig. 6B) provide insight into the most recent resurfacing event. These ages, 0.7 Ma (Fig. 5B) and 0.3 Ma (Fig. 6B), are the youngest crater retention ages observed in this study. The crater-size frequency distributions also exhibit the poorest fits to isochrons observed, which we attribute to several potential factors. We were conservative in our identification of craters and at the smallest sizes, pits and bulges along polygon troughs are similar to impact craters. If polygons develop as quickly as suggested (Mellon, 1997; Levy et al., 2009; Kreslavsky et al., 2011) and are being refreshed, that could lead to a deficit of meter-scale craters and a misinterpretation of decameter-size craters (which could be polygonalized).

Our best fits to isochrons (Fig. 5B; Fig. 6B; Table 1) are calculated using craters with diameters ≥ 8 m. If smaller craters were included, the best-fit estimates would be younger. Our count data suggest that the more poleward polygonal terrain (Fig. 6) has a factor of two younger crater retention age than the more equatorial polygonal terrain (Fig. 5), though both are very young.

Mid-latitude craters: The bulk of the craters in our study (11) are mid-latitude craters. These craters range in latitude from 20.8°S to 35.7°S and therefore bracket what has been interpreted as the equatorial extent of latitude-dependent mantling deposits, $\sim 30^\circ$ (Fig. 1). The two most equatorial craters in this group (Fig. 7; Fig. 8) exhibit no evidence of latitude-dependent mantling and have crater retention ages (Fig. 7B; Fig. 8B) comparable to the two older rayed craters discussed in section 3.1. Moving poleward, the next three craters in our study (Fig. 9; Fig. 10; Fig. 11) also do not exhibit any geomorphic evidence of latitude-dependent mantling. The crater in Fig. 9 is the smallest diameter crater (0.8 km) in our study. Crater retention ages are reported in Fig. 9B (5.8 Ma), Fig. 10B (62 Ma), and Fig. 11B (34.5 Ma). The un-mantled craters of Fig. 10 and Fig. 11 have crater retention ages that are a factor of ten older than what has been interpreted as the most recent ice age (Head et al., 2003).

The near-crater-rim area in Fig. 12A (27.4°S) exhibits prominent geomorphic evidence of remnant latitude-dependent mantle undergoing

degradation and erosion. Crater counts on the smooth mantle terrain indicate a crater retention age of 7.9 Ma for this surface (Fig. 12B). Erosion has revealed several depositional layers within the mantle, which is consistent with multiple episodes of mantle emplacement (Head et al., 2003; Schon et al., 2009a). Mantle destroying processes include sublimation of ice-content and eolian removal of the fines portion. Because individual sublimation pits are similar in size to small craters (Fig. 12A) it is not feasible to date the development of these features. Several factors may cause a net increase in the crater retention age for this unit relative to its age of emplacement. First, individual sublimation pits are difficult to distinguish from superposed impact craters and some superposed craters can actually be enlarged by sublimation, both factors leading to an artificially old crater retention age. Secondly, impact craters from the underlying pre-mantle surface might be exhumed by sublimation processes, introducing additional craters that would tend to produce older ages. Both of these factors are likely to more than offset any loss of craters due to mantle destruction, and thus we consider the 7.9 Ma crater retention age of the surface to be an upper limit for the mantle in this area.

Three young un-mantled craters (Fig. 13; Fig. 14; Fig. 15) near the dissected mantle boundary (Fig. 1) provide limits on the equatorial extent of recent mantling. The 7.3-km diameter crater in Fig. 13 is un-mantled and has a crater retention age of 2.2 Ma (Fig. 13B). The floor of this crater (Fig.

13A) has a characteristic hummocky texture that was noted by Tornabene et al. (2006) to occur in young craters. This texture has been interpreted to be the result of volatile-rich suevite that degassed rapidly in the terminal phases of the impact event (Boyce et al., 2011). Zumba crater (28.7°S, 226.9°E; Fig. 14A) also contains this texture (Tornabene et al., 2006; Hartmann et al., 2010). Zumba, a 2.9-km diameter crater occurring in Daedalia Planum on lava flows associated with Arsia Mons, lies in an area with no evidence of LDM deposits. Crater counts by Hartmann et al. (2010) suggest an age of 0.1 to 0.8 Ma. Our crater count data (Fig. 14B) indicate a best-fit age of 0.7 Ma for Zumba, consistent with the Hartmann et al. (2010) estimate. Because basaltic shergottites have ejection ages ranging from 20 Ma to 0.7 Ma and are from surface lava flows with crystallization ages of ~175 Ma and 330-475 Ma (Nyquist et al., 2001), Zumba is a plausible launch crater for some martian meteorites (e.g., Tornabene et al., 2006; Lang et al., 2009). An un-mantled 3.4-km diameter crater in Terra Cimmeria (Fig. 15) has a crater retention age of 0.9 Ma (Fig. 15B). This crater (29.5°S, 163.1°E) provides the temporal constraint that geologically-recent mantling deposits have not been emplaced here since formation of the crater.

Evidence of latitude-dependent mantling deposits and gullies are observed in association with a 1.8-km diameter crater (Fig. 16A) in Promethei Terra (32.2°S, 116.2°E). A triangular avoidance zone in the ejecta pattern (Fig. 16A) suggests this crater formed in an oblique impact (e.g.,

Gault and Wedekind, 1978). Mantling deposits within the crater are concentrated on the pole-facing wall and have a degraded morphology that is commonly associated with gully formation (Milliken et al., 2003; Christensen, 2003; Dickson and Head, 2009; Schon and Head, 2011a). In a survey of concentric crater fill deposits and younger latitude-dependent mantle related gullies (Dickson et al., 2011), similar isolated deposits have been observed to occur preferentially on pole-facing slopes at this latitude. Because mantling is not observed on the surrounding terrain, we interpret our crater count (Fig. 16B) to represent the formation age of the crater and a bound on the age of the degraded mantling and gullies within the crater, which must be younger (i.e., < 26.8 Ma).

Finally, Gasa crater (35.7°S, 129.4°E) provides an example of a young crater that post-dates regional latitude-dependent mantling (Schon and Head, 2011b). Secondaries from ~1.2 Ma Gasa have been used as a stratigraphic marker (Schon et al., 2009b) and the apex slopes of gully fans within Gasa have been compared to other mid-latitude gully deposits (Kolb et al., 2010). Gasa occurs within an 18-km diameter crater that is draped with latitude-dependent mantling deposits. Extensive evidence for the presence of a debris-covered glacier has been documented in the host crater, and melting of this ice by the Gasa impact is the likely source of meltwater for the development of gullies within Gasa (Schon and Head, 2011b). Therefore, although Gasa has gullies (Fig. 17A), it is a young crater (Fig. 17B) that

postdates emplacement but not degradation (gully formation) of latitude-dependent mantling in its region. Crater rays from Gasa are observed on the latitude-dependent mantle (Schon et al., 2009b).

Interpretations of LDM Chronology

Obliquity-driven climate change has long been recognized as an important feature of the Amazonian (e.g., Ward, 1973; Sagan et al., 1973; Soderblom et al., 1973; Toon et al., 1980; Touma and Wisdom, 1993). Recognition of the latitude-dependent mantle and its youthfulness (Kreslavsky and Head, 2000; Mustard et al., 2001; Kreslavsky and Head, 2002) led to the interpretation of a recent ice age during a period of enhanced obliquity variation from 2.1 Ma to 0.4 Ma (Head et al., 2003). With new sub-meter resolution image data, crater counts on homogenized surfaces can be used to constrain the history of this deposit. We present the isochron fits to our crater counts in Table 1. Of course using small craters and areas for dating leads to some uncertainties (Hartmann, 2005; Hartmann, 2007; Hartmann et al., 2010) and we thus do not use these values for specific individual age constraints, but rather we base our interpretations and conclusions on multiple crater counts and factor-of-several to factor-of-ten differences in crater retention ages.

In our interpretation, crater retention ages < 1 Ma (Fig. 5; Fig. 6; cf. Kostama et al., 2006; Levy et al., 2009; Kreslavsky et al., 2011), in

conjunction with the pervasive mantling and polygonalization of decameter and larger craters, are consistent with the emplacement of ice-rich latitude-dependent mantling during the most recent ice age, 2.1 – 0.4 Ma (Fig. 18; Fig. 19). Crater retention ages of < 1 Ma and the timescale of polygon formation suggest that thermal cycling could form polygons under current conditions (e.g., Korteniemi and Kreslavsky, 2011). The absence of rayed craters from these latitudes also supports our interpretation of geologically recent mantling events.

Our observations indicate that the current equatorial margin of remnant latitude-dependent mantle is variable, as might be expected based on regional depositional heterogeneity, weather and climate patterns, and preservation potential (e.g., Costard et al., 2001; Madeleine et al., 2009; Morgan et al., 2010). The most equatorial mantle with a crater retention age of ~7.9 Ma (Fig. 12) is at least a factor of several older than the high latitude mantle terrain (Fig. 5; Fig. 6). This suggests that the mantle was not emplaced synchronously across the mid- and high- latitudes. The presence of ~1.2 Ma Gasa crater rays and secondaries on mid-latitude mantle surfaces (Schon et al., 2009b; Schon and Head, 2011b) also indicates that mid-latitude LDM is older than high latitude mantle. Although substrate is important for recognizing rays on Mars (Tornabene et al., 2006), our data show that crater rays in the equatorial region can persist for tens of millions of years. In

contrast, other rayed craters similar to Gasa, are not observed superposed on the LDM, which further supports the young age of the mantle.

What was the source region of the ice for the ice-rich latitude-dependent mantle? The polar caps have been suggested as a potential source, but climate models indicate that an equatorial ice source (formed at high obliquity) is necessary to precipitate the mantling deposits (Levrard et al., 2004). With the exhaustion of an equatorial ice source, the equatorial extent of the LDM would become unstable and simulations of Levrard et al. (2004) indicate that surface ice would be re-deposited poleward. Our data and observations of older dissected mantle relative to younger high latitude LDM provide geological support for this scenario of poleward redistribution during the waning of an ice age. Was the entirety of currently observed latitude-dependent mantling deposits (Fig. 1) formed during the last ice age (Head et al., 2003), a period of enhanced obliquity ~2.1 to 0.4 Ma (Fig. 18)? Or, could mid-latitude portions of the mantle be remnant from the transition to lower mean obliquity that occurred ~5 Ma? Given uncertainties in the crater retention ages, our observations are consistent with either timing scenario. The large variations of Mars obliquity and the substantial probability of high obliquity periods over geologic time suggest that the current latitude-dependent mantle may be only the most recent manifestation of a cyclic process (Laskar et al., 2004; Levrard et al., 2004).

Conclusions

Chronologies based on small craters must be interpreted carefully because uncertainties are potentially greater than with larger craters and older surfaces (Hartmann, 2005; Hartmann, 2007; Hartmann et al., 2010). Our analysis of more than 60,000 superposed craters on the near-rim deposits of 16 young craters supports the following conclusions:

- (1) Detailed counts of small craters on geologically young surfaces have size-frequency distributions consistent with the isochrons of Hartmann (2005).
- (2) Rayed craters in the equatorial regions, e.g., Thila (Fig. 2) and Dilly (Fig. 4), can retain their rays for a period likely to be tens of millions of years.
- (3) The ages of rayed craters (Table 1, Hartmann et al., 2010) are consistent with the isochron system and exceed the formation intervals expected (Hartmann et al., 2010; cf. McEwen et al., 2005).
- (4) The ice-rich latitude-dependent mantle is geologically young. The LDM has crater-retention ages approximately a factor of 10 less than the timescale for crater-ray retention at low latitudes (Table 1). The youngest LDM is found at high latitudes where polygonal patterned ground is pervasive and is $\ll 1$ Ma (Levy et al., 2009; Kreslavsky et al., 2011).
- (5) Remnant latitude-dependent mantling (e.g., Fig. 12) is limited to the dissected mantle region (Fig. 1) and is older than high latitude polygonally patterned ground by a factor of several or more.

(6) The age and latitudinal-age trend of the LDM (younger at higher latitudes) is consistent with suggested control by obliquity variations (Fig. 18; Head et al., 2003; Schorghofer, 2007).

(7) The precise extent of remnant LDM is likely to be related to regional depositional heterogeneity (e.g., Levrard et al., 2004) and local preservation potential.

(8) Ice-rich LDM was deposited in the region of polygonally-patterned ground during the last ice age (2.1-0.4 Ma, Head et al., 2003). Polygon development due to thermal cycling of previously emplaced ice is likely to be ongoing (Mellon, 1997; Levy et al., 2009; Korteniemi and Kreslavsky, 2011).

(9) Remnant latitude-dependent mantle (e.g., Fig. 12) in the dissected region (Fig. 1) dates from either earlier in the last ice age (~2.1 Ma) or from the transition from higher mean obliquity at ~5 Ma (Fig. 18).

Acknowledgments

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Table 1:

Fig. (Name)	Longitude (E)	Latitude	Diam. (km)	HiRISE Observation	Age (Ma) [D > 8 m]	# of craters counted	Count Area (km ²)
2 (Thila) ¹	155.6	18.1	5.4	PSP_009346_1985	23.1	2790	4.2
3 (Naryn) ²	123.3	14.9	3.9	PSP_002570_1950	2.1	3169	4.5
4 (Dilly) ³	157.2	13.3	2.1	PSP_010203_1935	34.4	6675	18.9
7	150.6	-20.8	7.3	PSP_010032_1590	31.5	12482	3.8
8	94.8	-23.9	4.3	PSP_008030_1560	18.1	4355	6.0
9	49.4	-24.9	0.8	PSP_009983_1550	5.8	860	6.5
10	115.7	-25.8	6.9	PSP_007436_1540	62	10710	3.3
11	127.1	-25.9	2.9	PSP_008543_1540	34.5	6749	8.9
12	59.1	-27.4	6.5	ESP_014400_1525	7.9	865	12.6
13	108.9	-28	7.3	PSP_004008_1520	2.2	1451	7.0
14 (Zumba) ⁴	226.9	-28.7	2.9	PSP_003608_1510	0.7	1197	46.0
15	163.1	-29.5	3.4	ESP_020752_1500	0.9	5408	4.5
16	116.2	-32.2	1.8	ESP_020701_1475	26.8	4834	2.9
17 (Gasa) ⁵	129.4	-35.7	7.0	PSP_004060_1440	1.2	289	6.8
5	46.4	-55.6	2.2	PSP_007030_1240	0.7	129	5.0
6	345.4	-77.9	1.4	ESP_013968_1020	0.3	332	15.1

¹Tornabene et al. (2006). ²Tornabene et al. (2006); Hartmann et al. (2010). ³Tornabene et al. (2006). ⁴Tornabene et al. (2006); Hartmann et al. (2010). ⁵Schon et al. (2009b); Kolb et al. (2010); Schon and Head (2011b). ⁶Count areas are variable due to factors including the extent of near-rim ejecta deposits, pitted areas that were avoided, and local topography that could make recognition of small craters difficult.

Fig.s

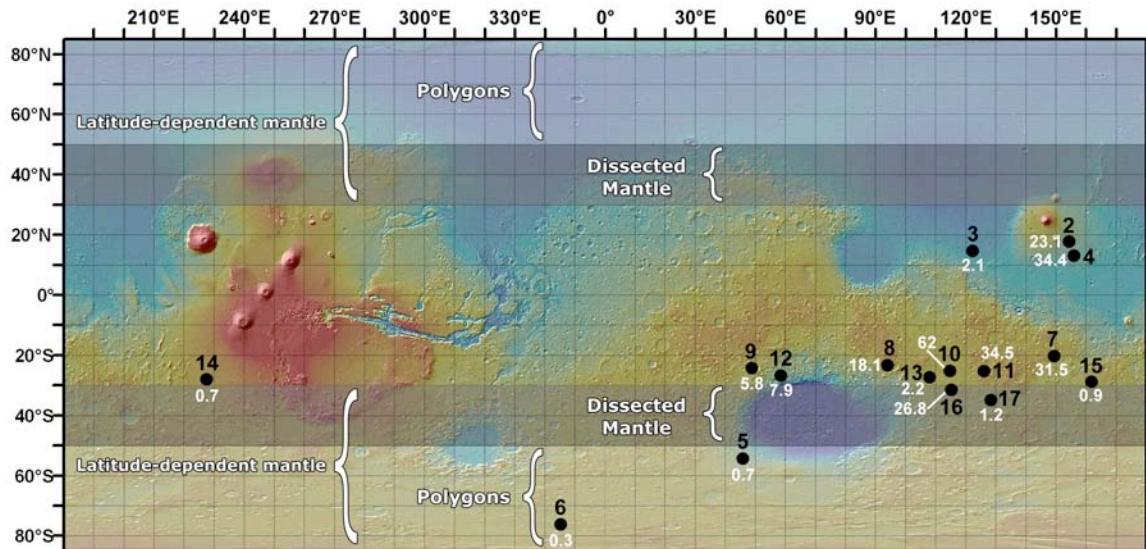


Fig. 1: Layers of ice-rich latitude-dependent mantling are found in both hemispheres (Head et al., 2003; 2011). Poleward of $\sim 60^\circ$ the terrain is dominated by patterned ground interpreted as contraction crack polygons. Between approximately 30° and 60° , where gullies and viscous flow features are commonly observed, the mantle is pitted, degraded, and partially removed. Dissected mantle refers to the remnant deposit of the former ice-rich dust layers. Equatorial rayed craters Thila (2), Naryn (3), and Dilly (4) were identified by Tornabene et al. (2006). Additional young craters in this study occur in the polygonal terrain (5 and 6) and bracket the equatorial extent of the mantle in the southern hemisphere ($\sim 30^\circ\text{S}$). Black numbers correspond to the associated Fig.. White numbers are crater-retention ages (Table 1). Background topography is from MOLA (red is high, blue is low).

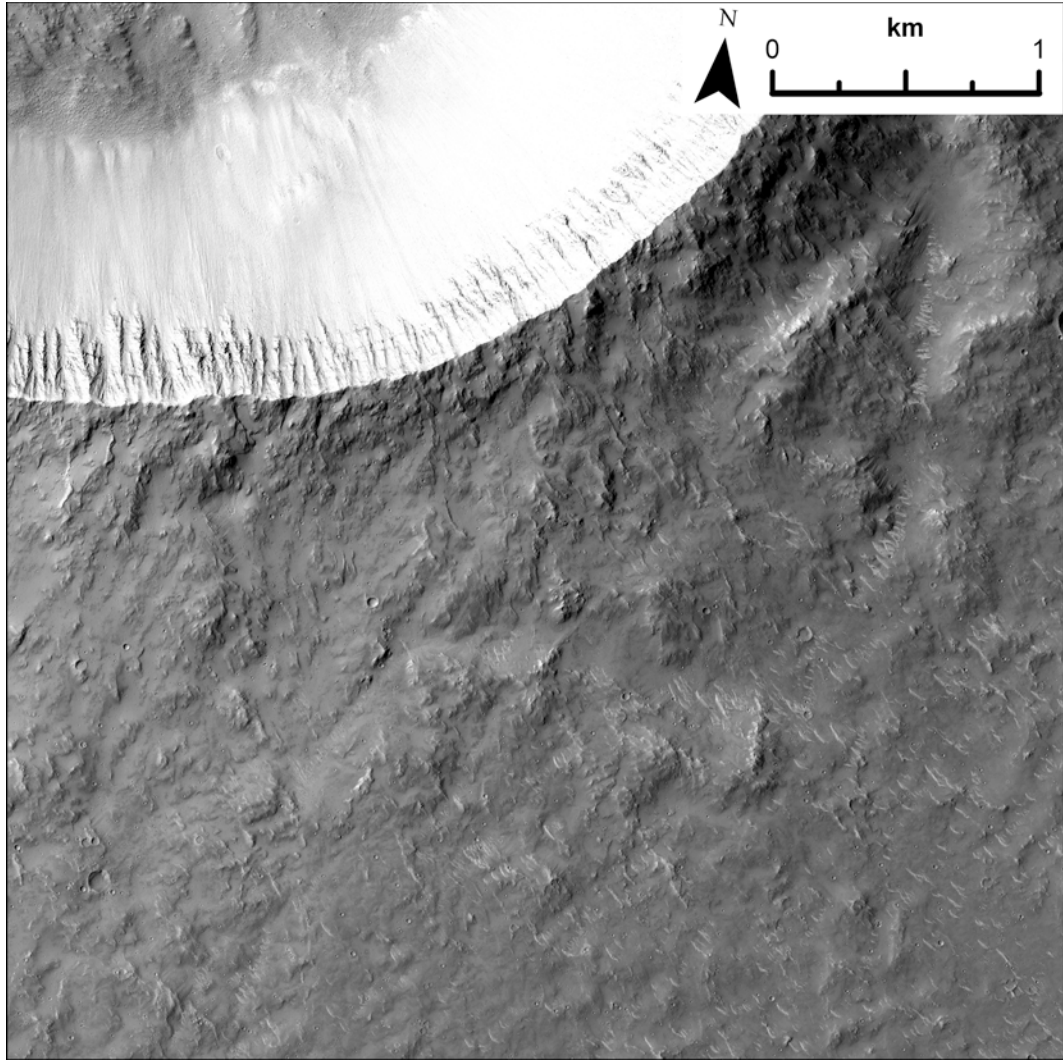


Fig. 2A: Thila crater (18.1°N, 155.6°E) is a 5.4-km diameter rayed crater identified by Tornabene et al. (2006) in Elysium Planitia. Portion of HiRISE: PSP_009346_1985.

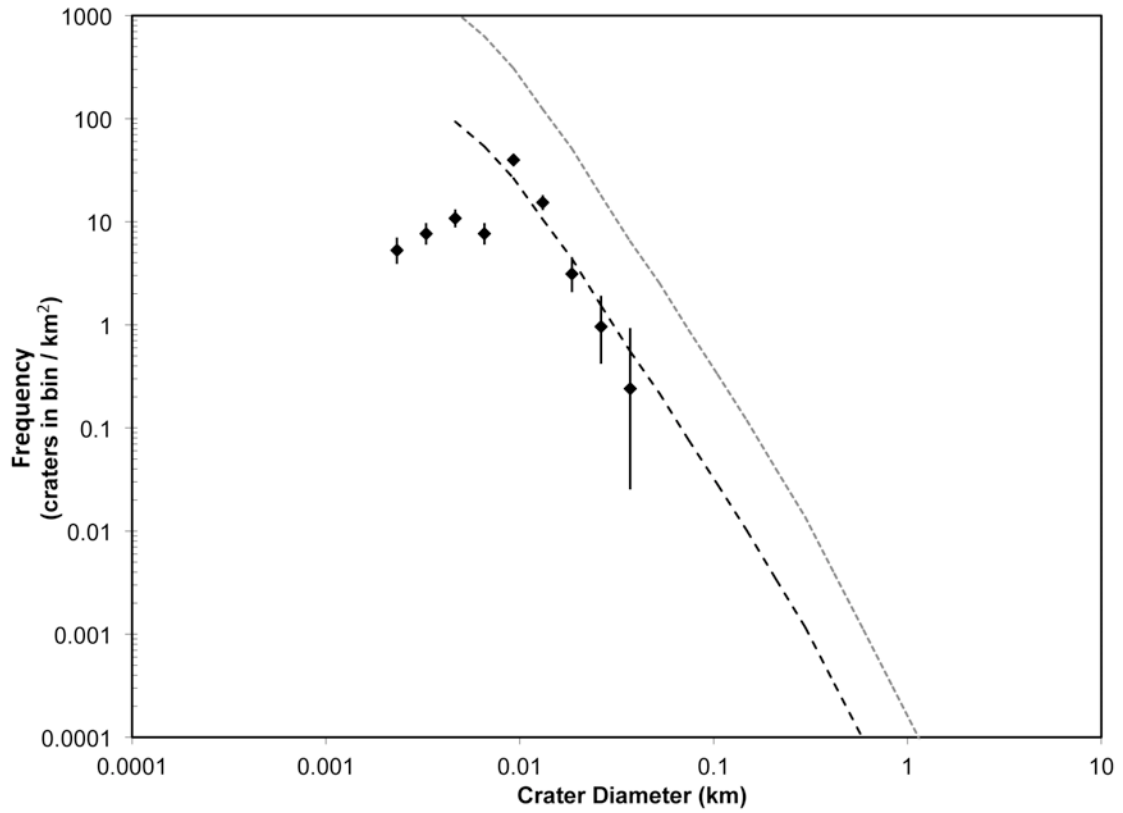


Fig. 2B: A crater count revealed 2,790 craters on 4.2 km² of near-rim deposits surrounding Thila crater. Isochrons of Hartmann (2005) indicate a best-fit age of 23.1 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

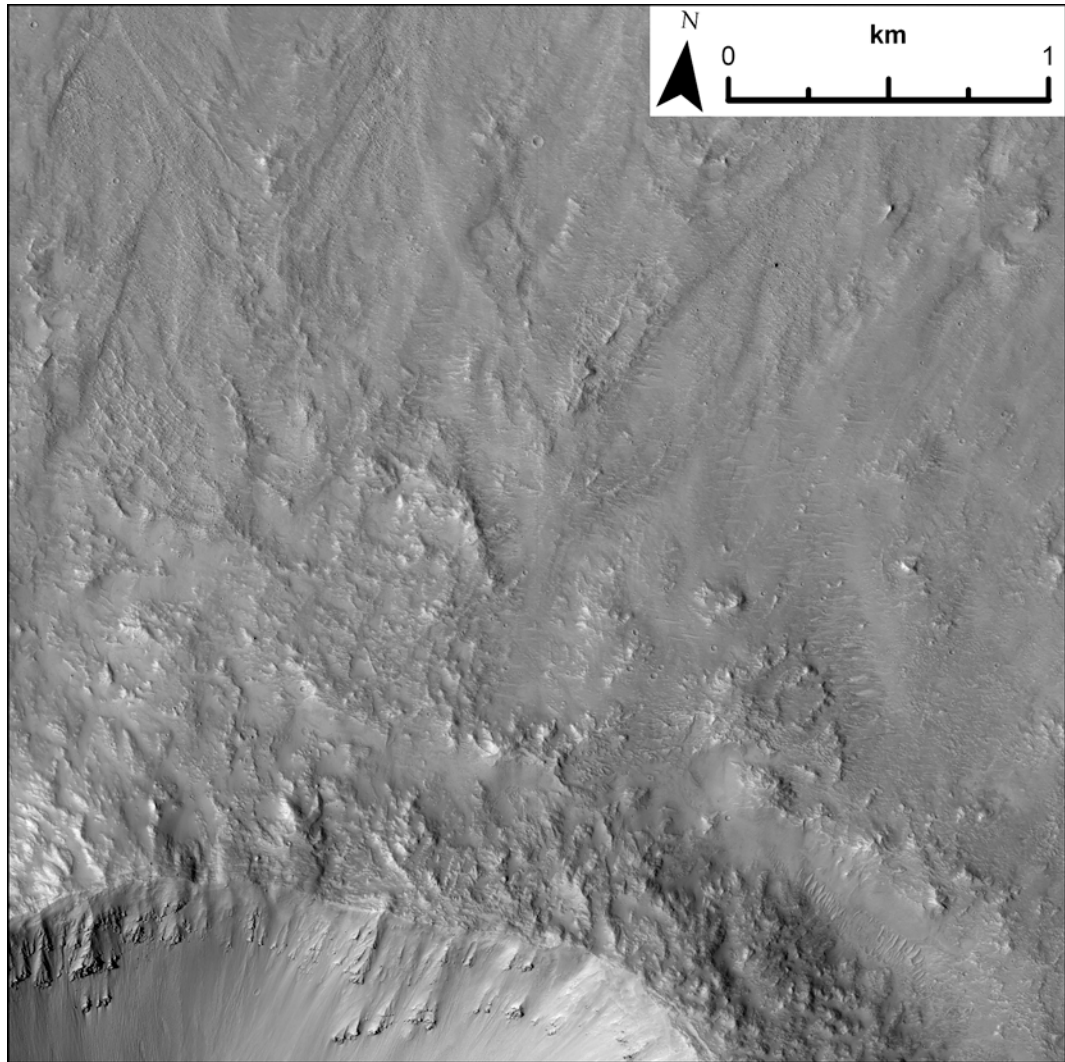


Fig. 3A: Naryn crater (14.9°N, 123.3°E) is a 3.9-km diameter rayed crater identified by Tornabene et al. (2006) in Elysium Planitia. Portion of HiRISE: PSP_002570_1950.

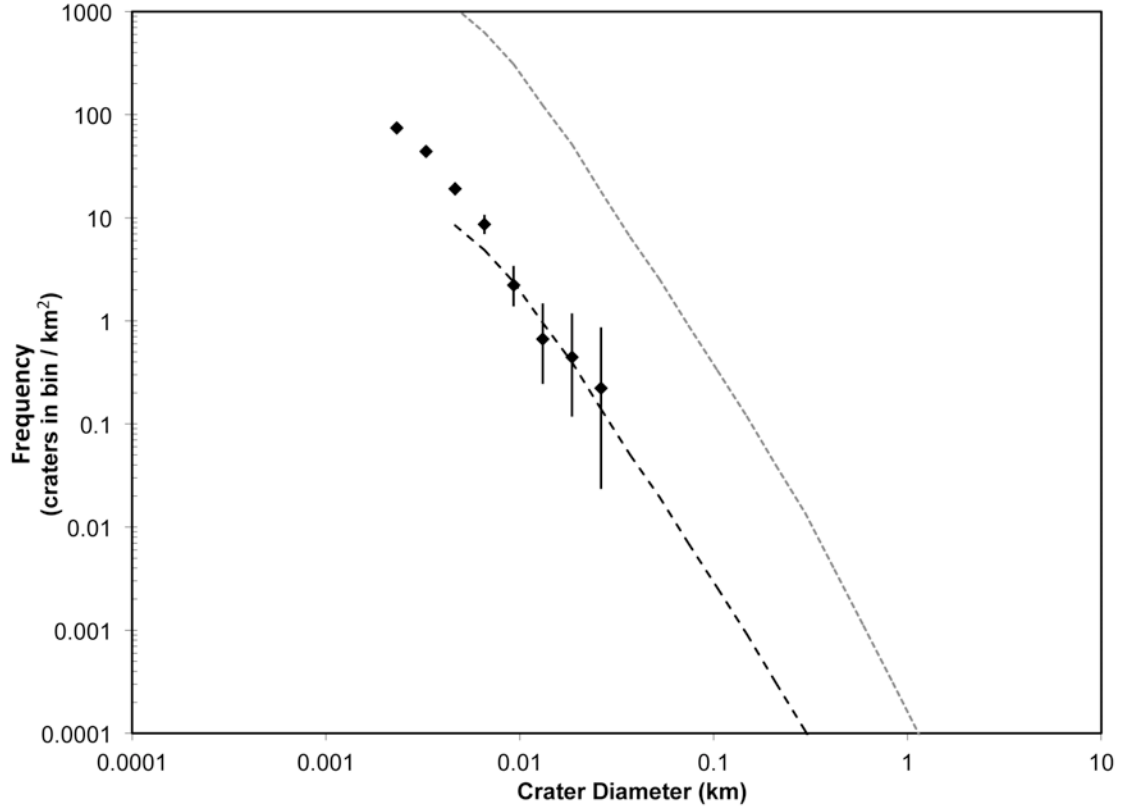


Fig. 3B: A crater count revealed 3,169 craters on 4.5 km² of near-rim deposits surrounding Naryn crater. Isochrons of Hartmann (2005) indicate a best-fit age of 2.1 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

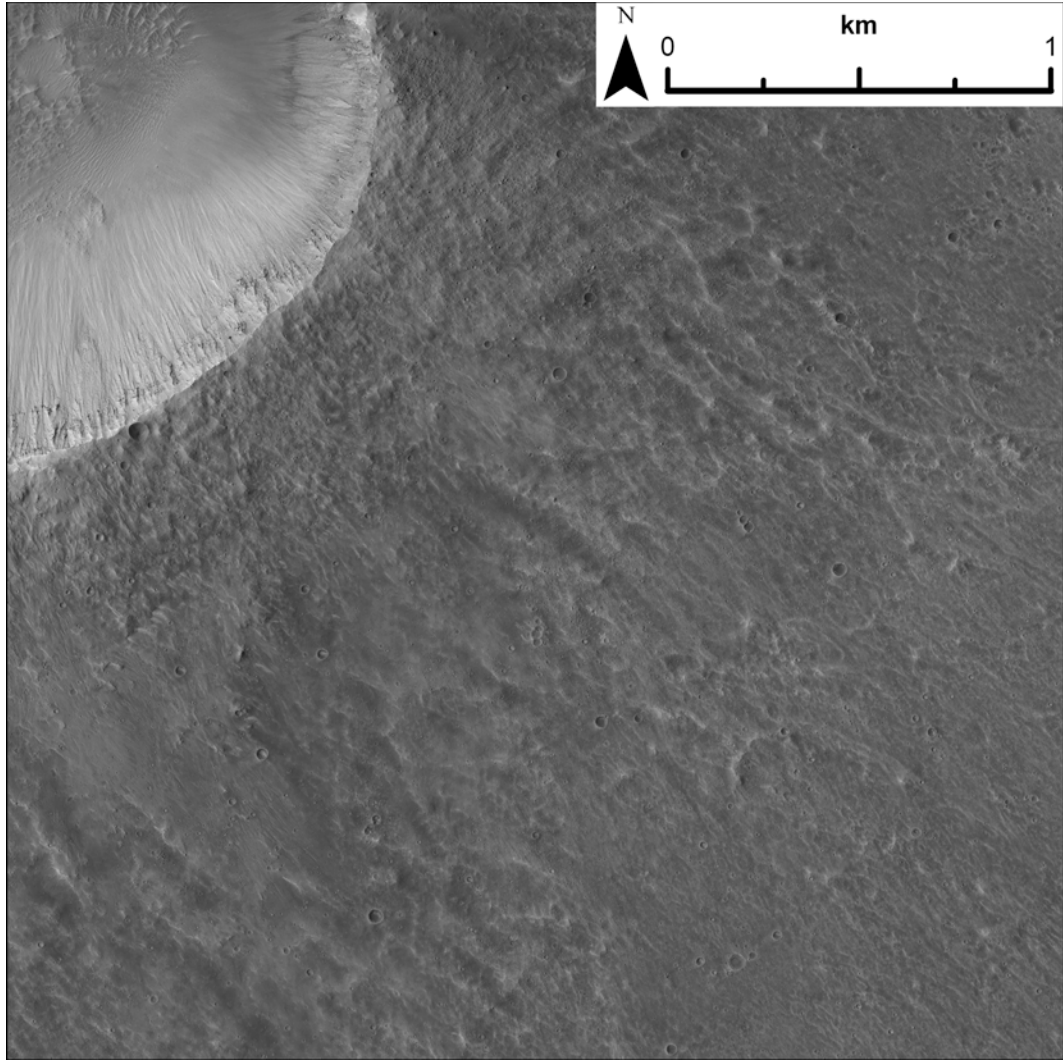


Fig. 4A: Dilly crater (13.3°N, 157.2°E) is 2.1-km diameter rayed crater identified by Tornabene et al. (2006) in Elysium Planitia. Portion of HiRISE: PSP_010203_1935.

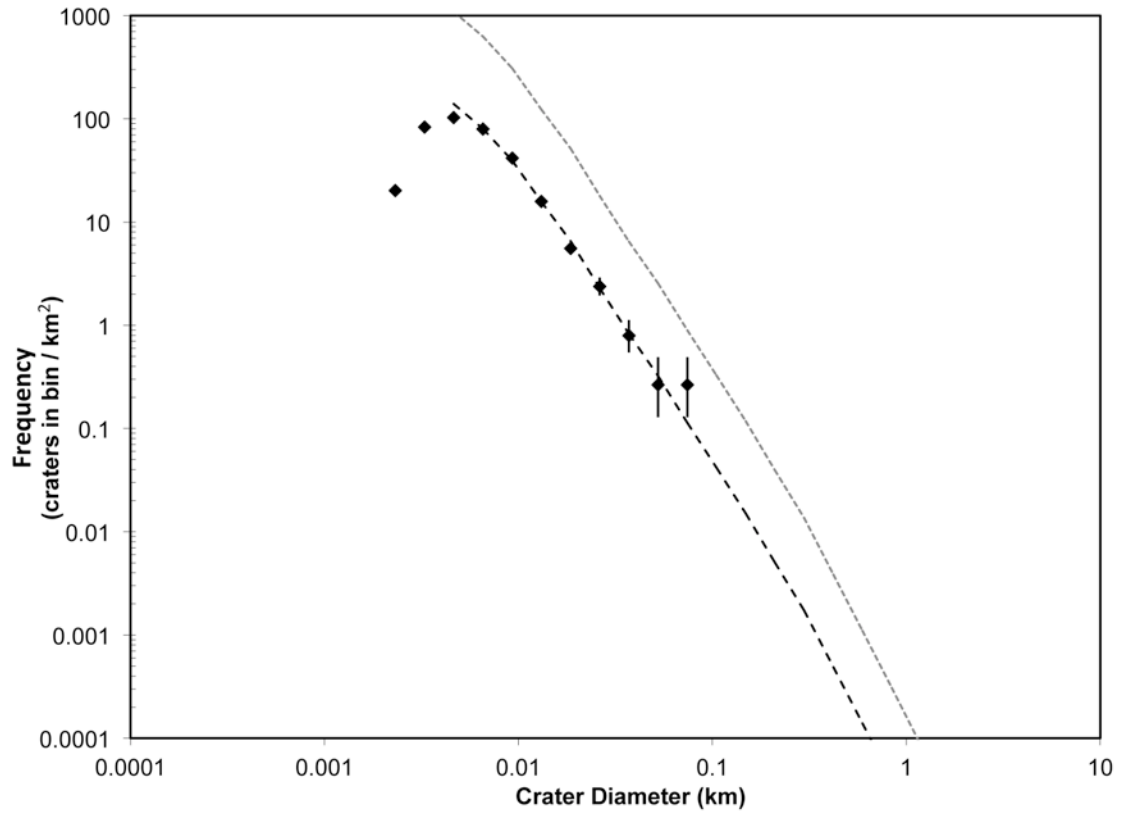


Fig. 4B: A crater count revealed 6,675 craters on 18.9 km² of near-rim deposits surrounding Dilly crater. Isochrons of Hartmann (2005) indicate a best-fit age of 34.4 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

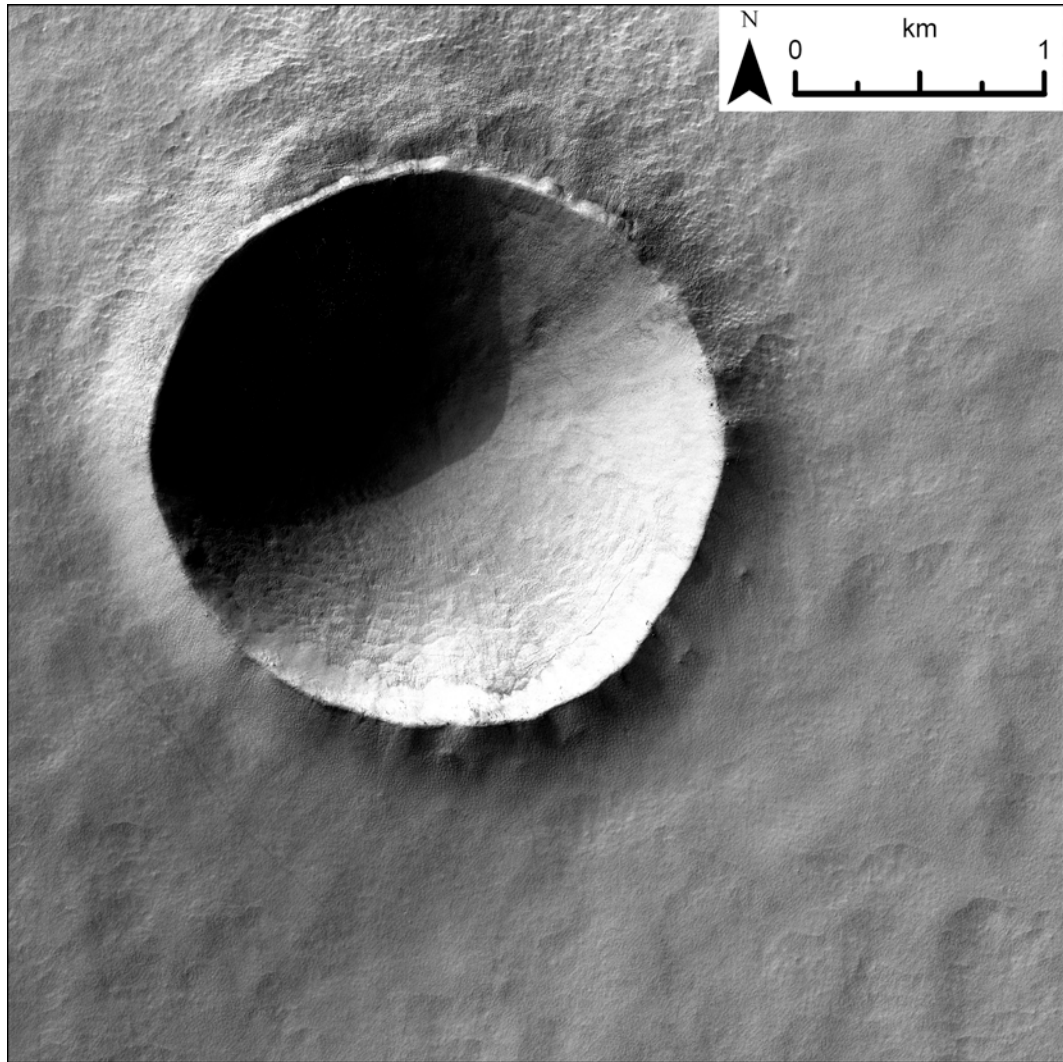


Fig. 5A: An unnamed 2.2-km diameter crater is found near the boundary of Noachis Terra and Hellas Planitia (55.6°S, 46.4°E). Polygonally patterned ground is found pervasively on the crater floor, wall, and rim, and on surrounding terrain. Portion of HiRISE: PSP_007030_1240.

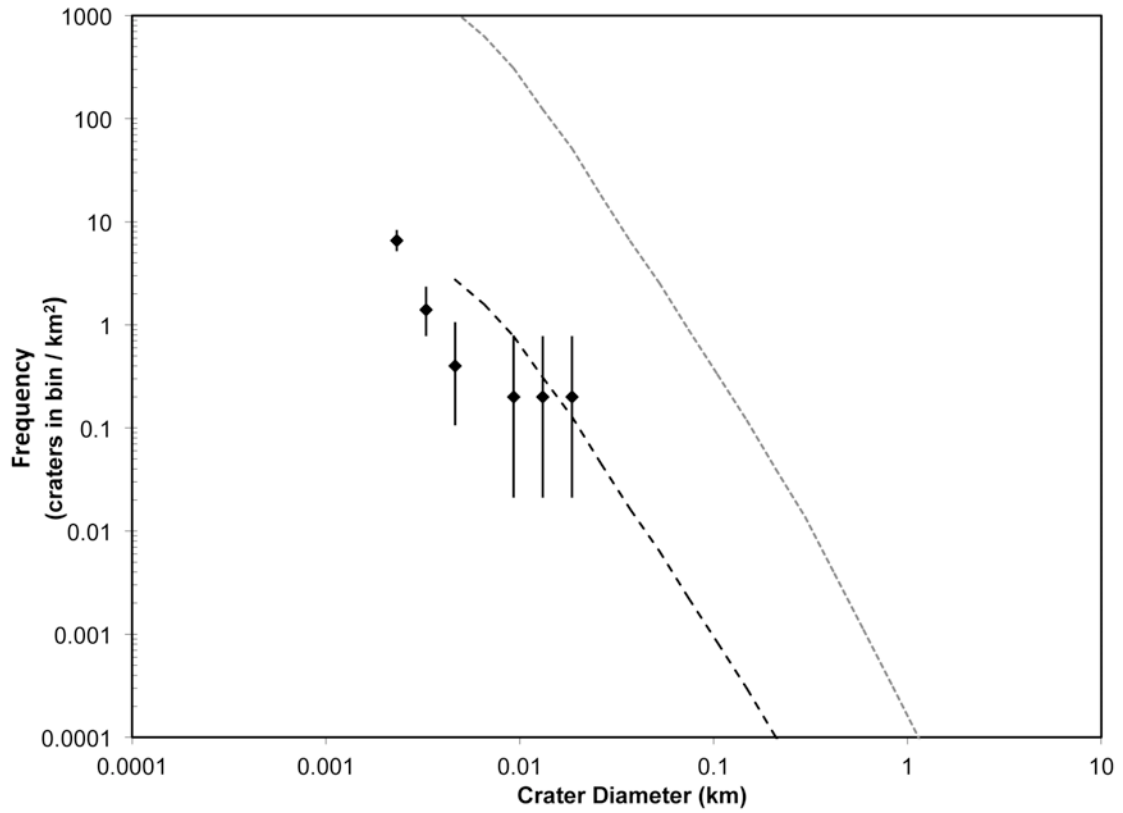


Fig. 5B: A crater count revealed 129 craters on 5.0 km² of near-rim deposits surrounding the unnamed crater (55.6°S, 46.4°E). Isochrons of Hartmann (2005) indicate a best-fit age of 0.7 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

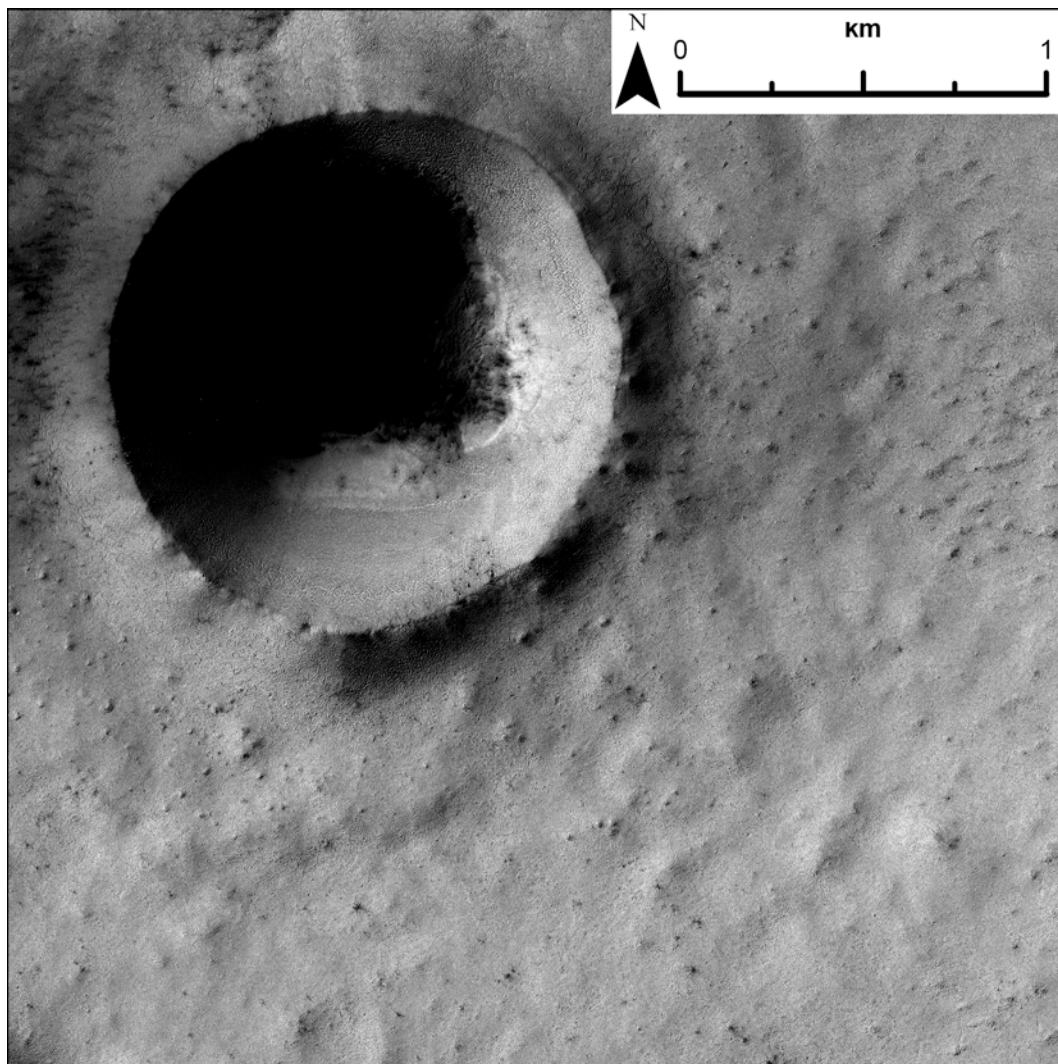


Fig. 6A: An unnamed 1.4-km diameter crater is found in Sisyphi Planum (77.9°S, 345.4°E). Polygonally patterned ground is found pervasively on the crater floor, wall, and rim, and on surrounding terrain. Portion of HiRISE: ESP_013968_1020.

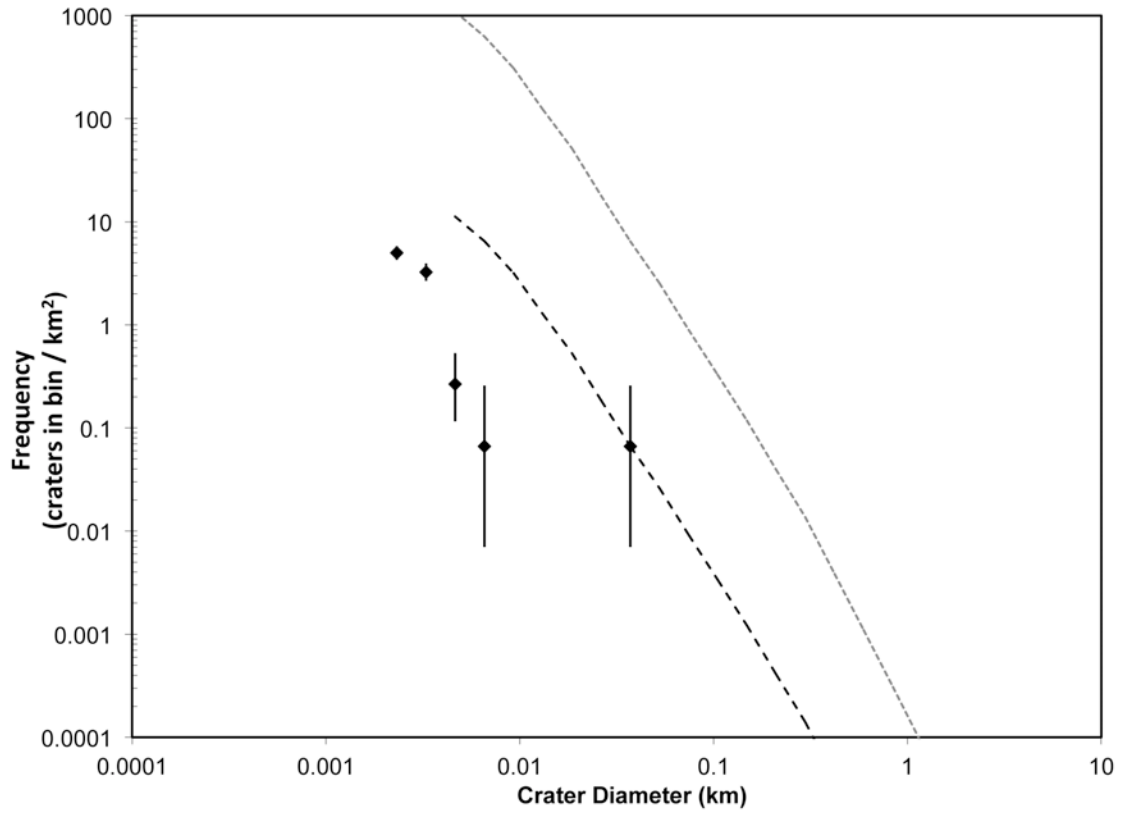


Fig. 6B: A crater count revealed 332 craters on 15.1 km² of near-rim deposits surrounding the unnamed crater (77.9°S, 345.4°E). Isochrons of Hartmann (2005) indicate a best-fit age of 0.3 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

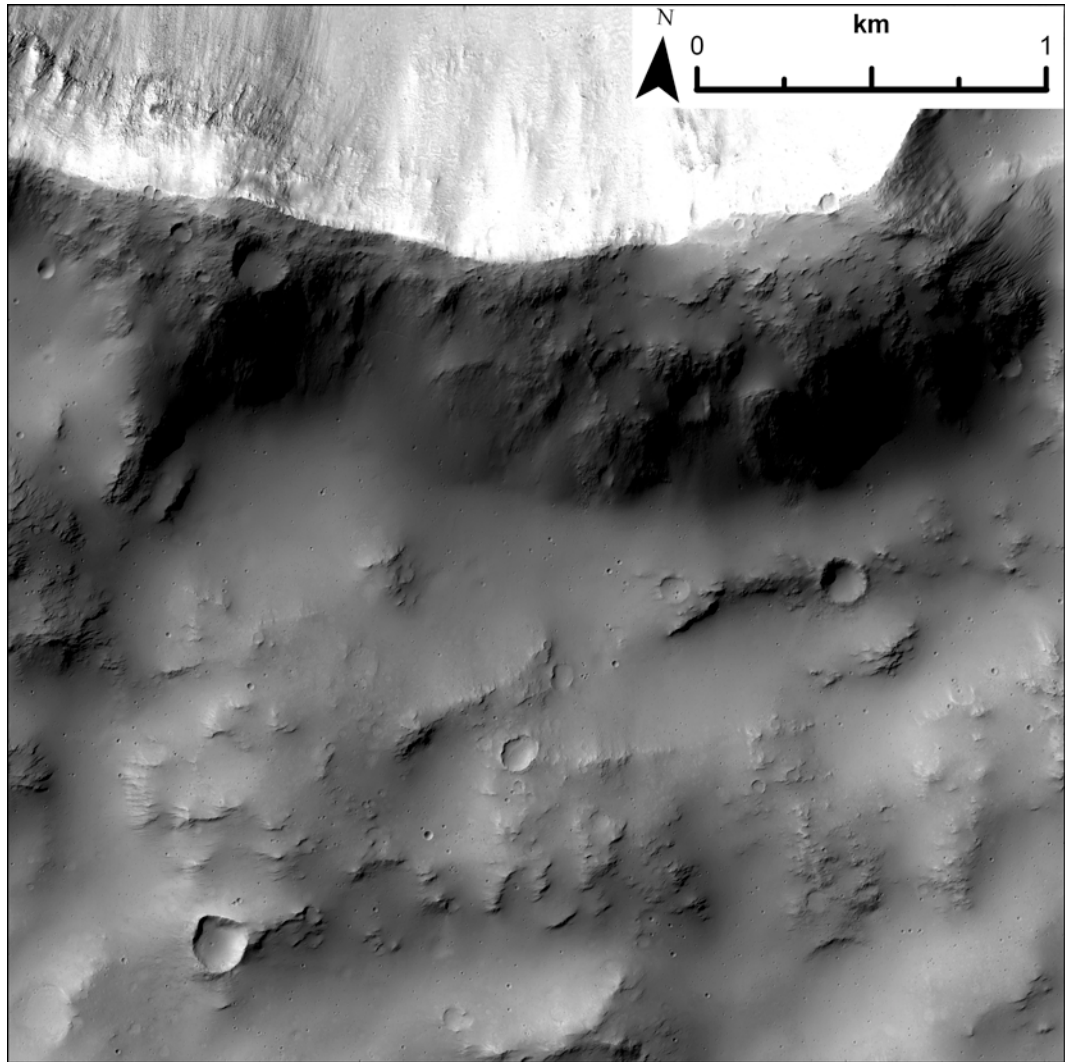


Fig. 7A: An unnamed 7.3-km diameter crater is found in Terra Cimmeria (20.8°S, 150.6°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_010032_1590.

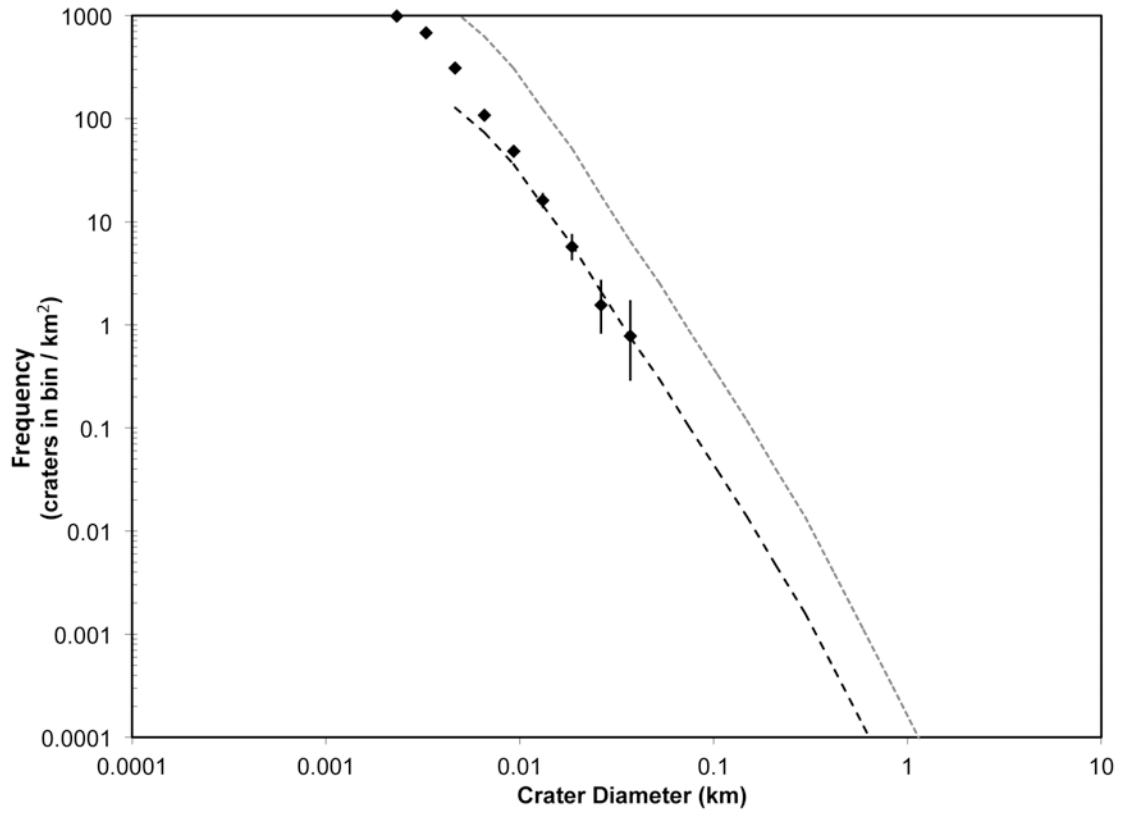


Fig. 7B: A crater count revealed 12,482 craters on 3.8 km² of near-rim deposits surrounding the unnamed crater (20.8°S, 150.6°E). Isochrons of Hartmann (2005) indicate a best-fit age of 31.5 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

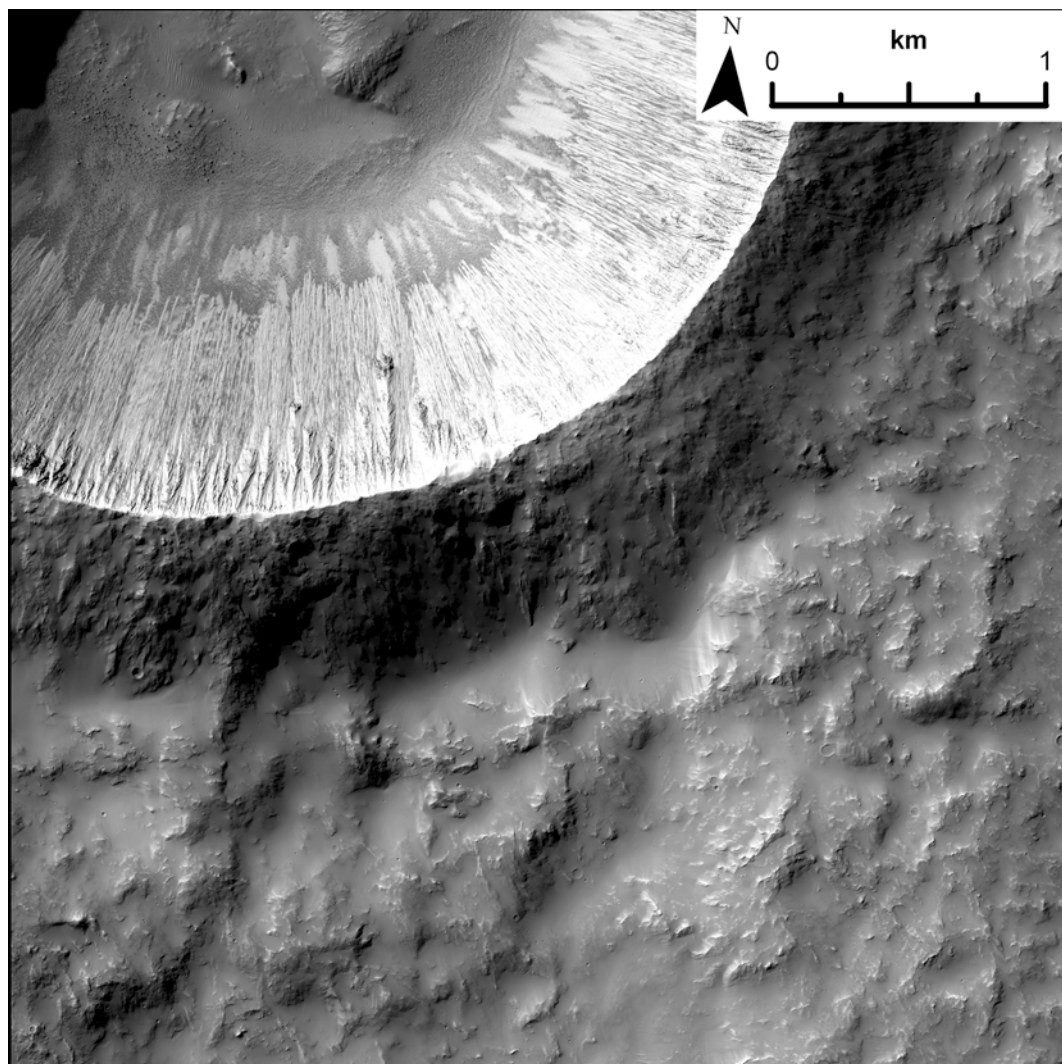


Fig. 8A: An unnamed 4.3-km diameter crater is found in Tyrrhenia Terra (23.9°S, 94.8°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_008030_1560.

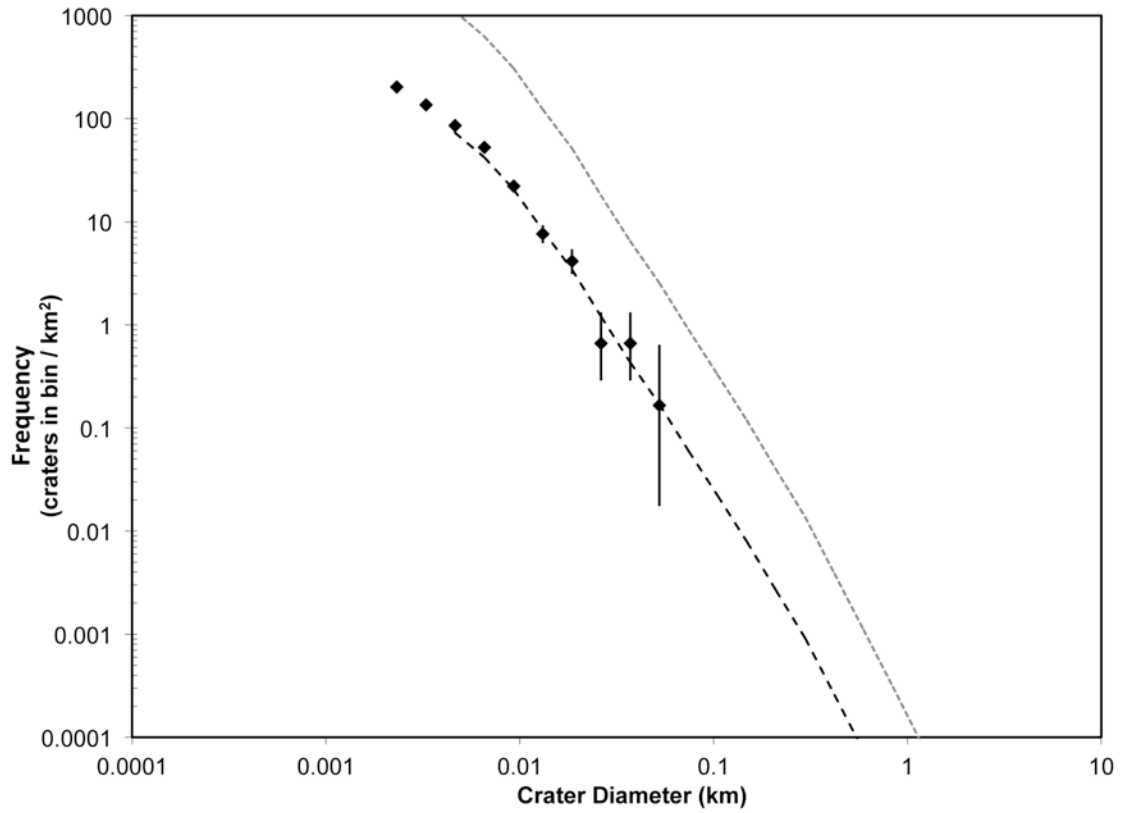


Fig. 8B: A crater count revealed 4,355 craters on 6.0 km² of near-rim deposits surrounding the unnamed crater (23.9°S, 94.8°E). Isochrons of Hartmann (2005) indicate a best-fit age of 18.1 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

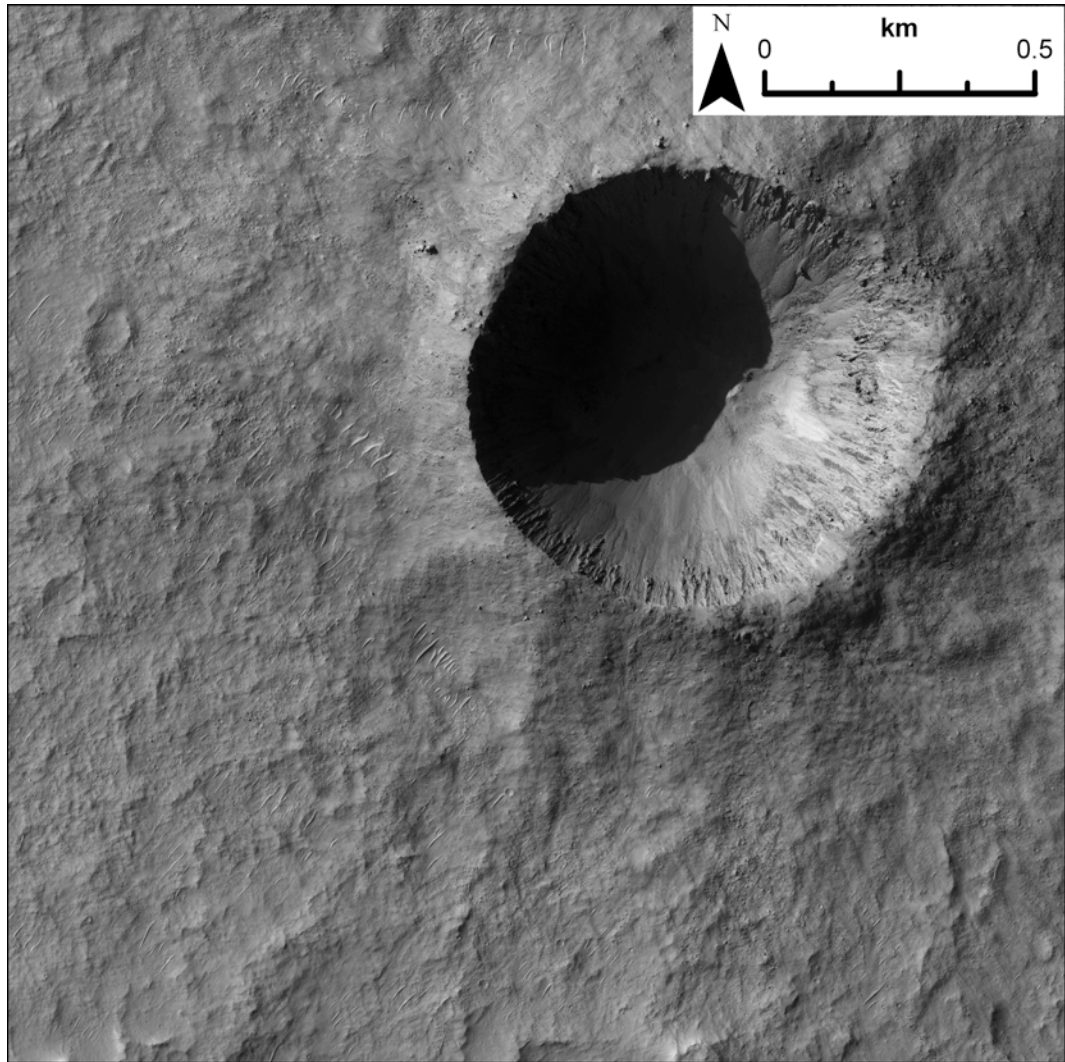


Fig. 9A: An unnamed 0.8-km diameter crater is found in Terra Sabaea (24.9°S, 49.4°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_009983_1550.

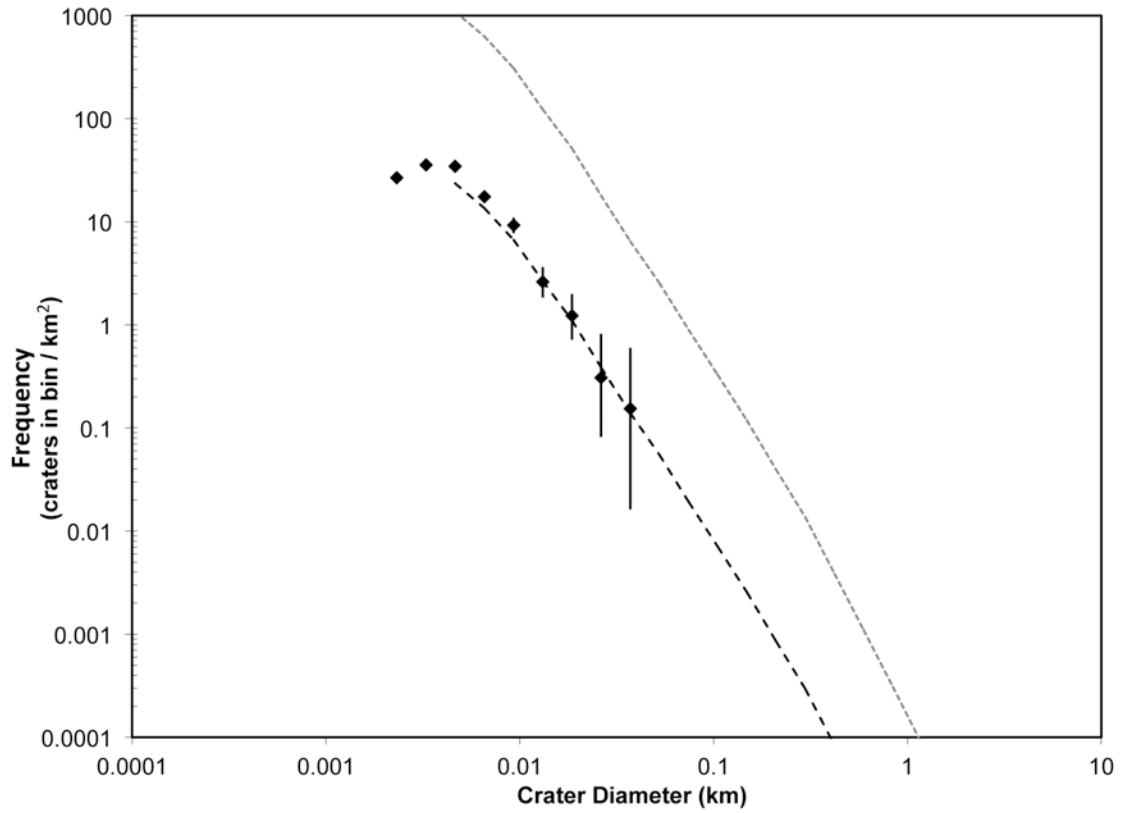


Fig. 9B: A crater count revealed 860 craters on 6.5 km² of near-rim deposits surrounding the unnamed crater (24.9°S, 49.4°E). Isochrons of Hartmann (2005) indicate a best-fit age of 5.8 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

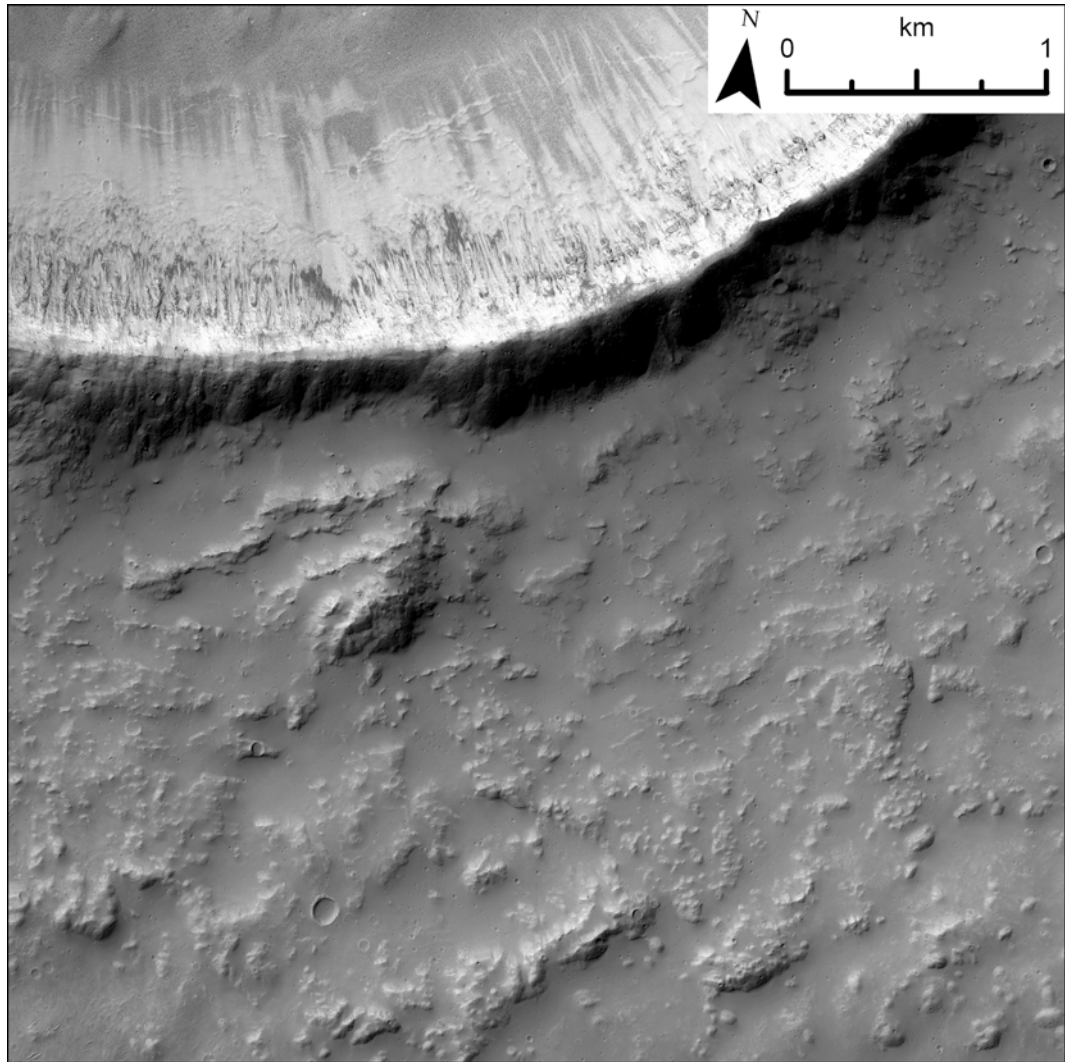


Fig. 10A: An unnamed 6.9-km diameter crater is found in Hesperia Planum (25.8°S, 115.7°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_007436_1540.

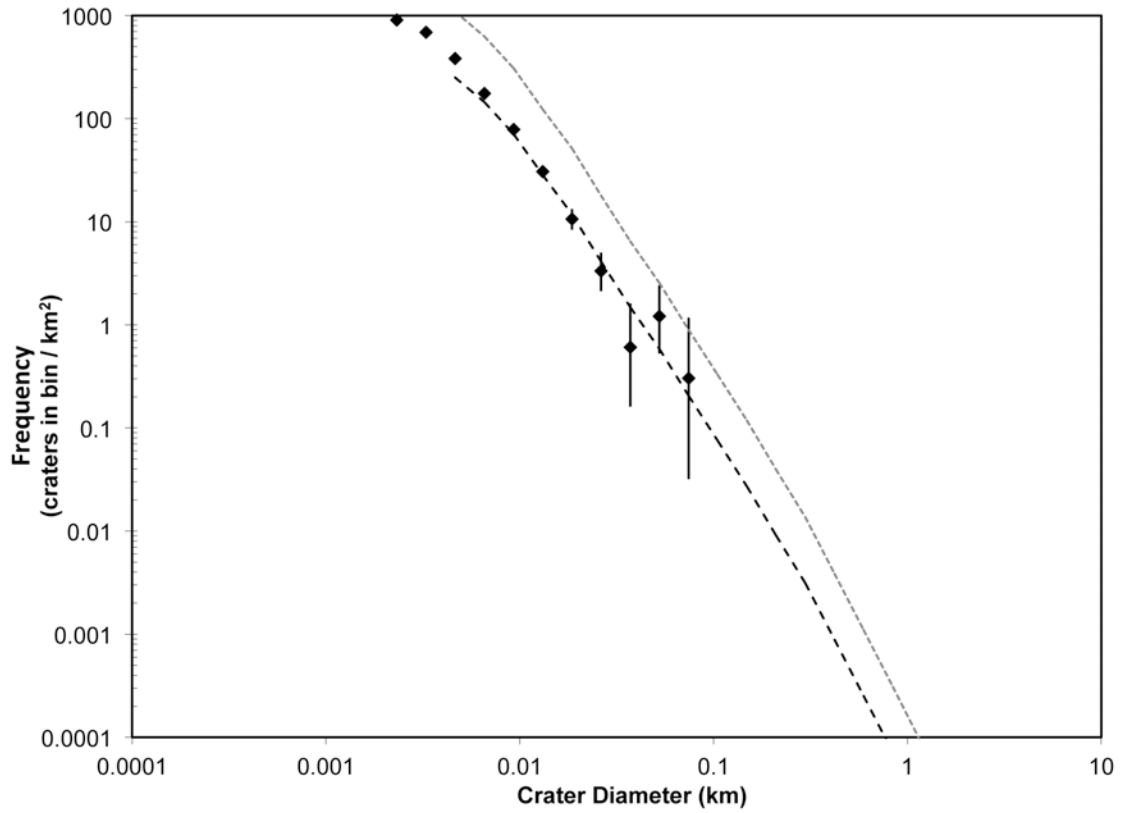


Fig. 10B: A crater count revealed 10,710 craters on 3.3 km² of near-rim deposits surrounding the unnamed crater (25.8°S, 115.7°E). Isochrons of Hartmann (2005) indicate a best-fit age of 62.0 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

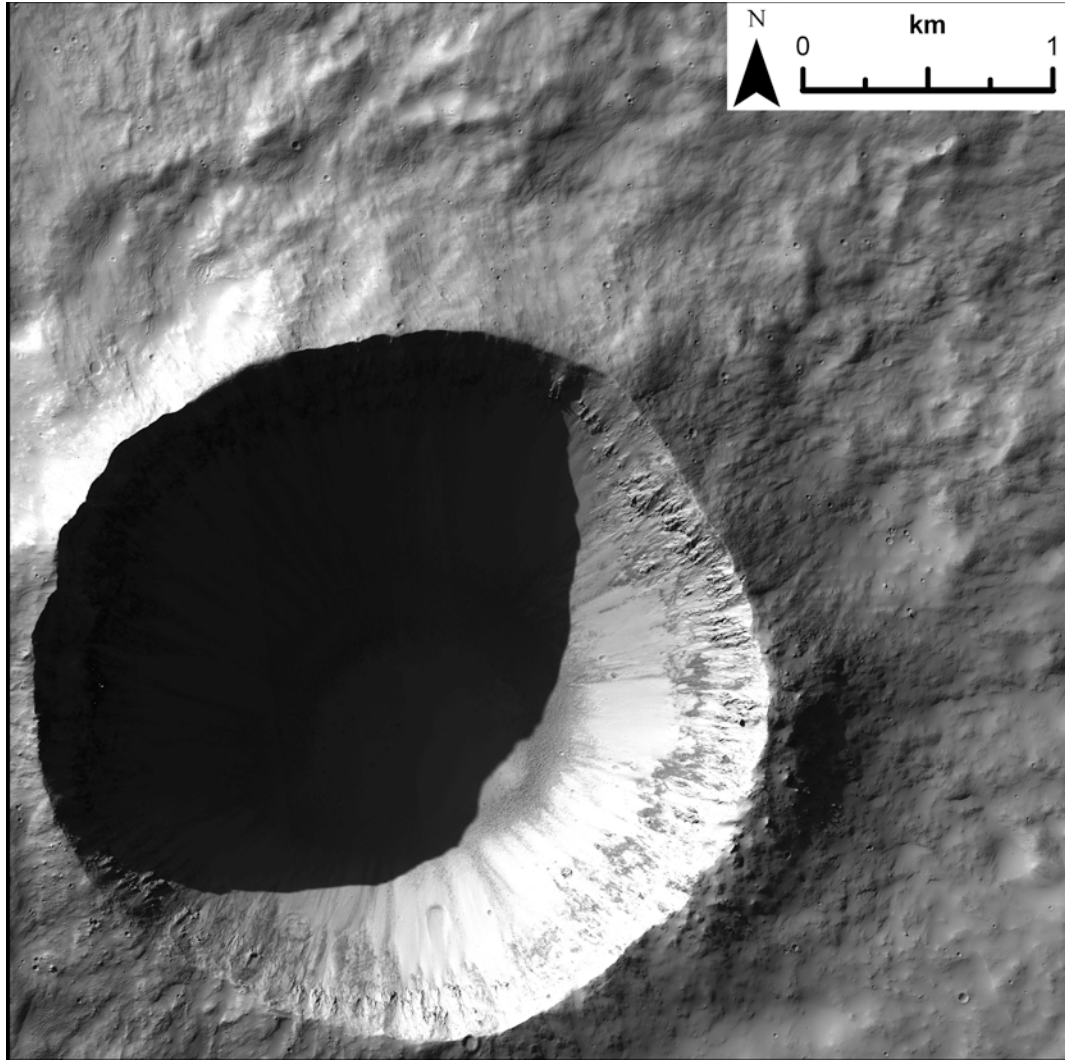


Fig. 11A: An unnamed 2.9-km diameter crater is found in Terra Cimmeria (25.9°S, 127.1°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_008543_1540.

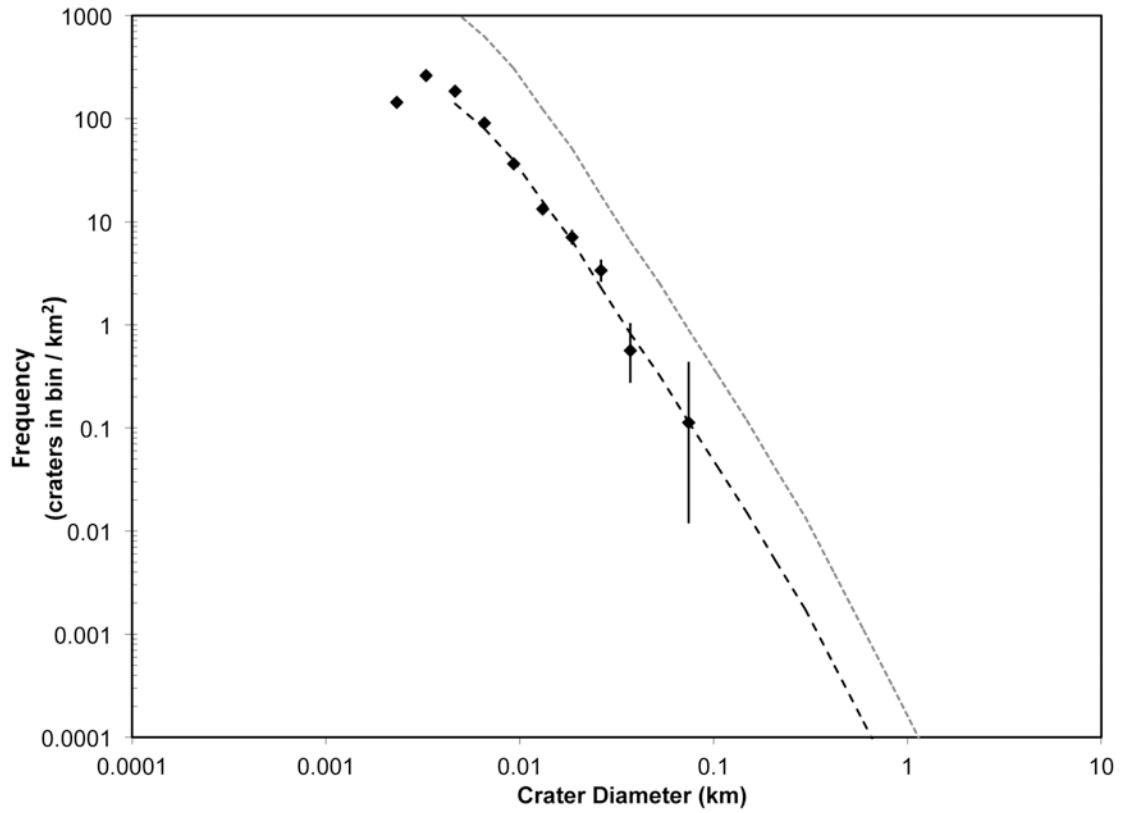


Fig. 11B: A crater count revealed 6,749 craters on 8.9 km² of near-rim deposits surrounding the unnamed crater (25.9°S, 127.1°E). Isochrons of Hartmann (2005) indicate a best-fit age of 34.5 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

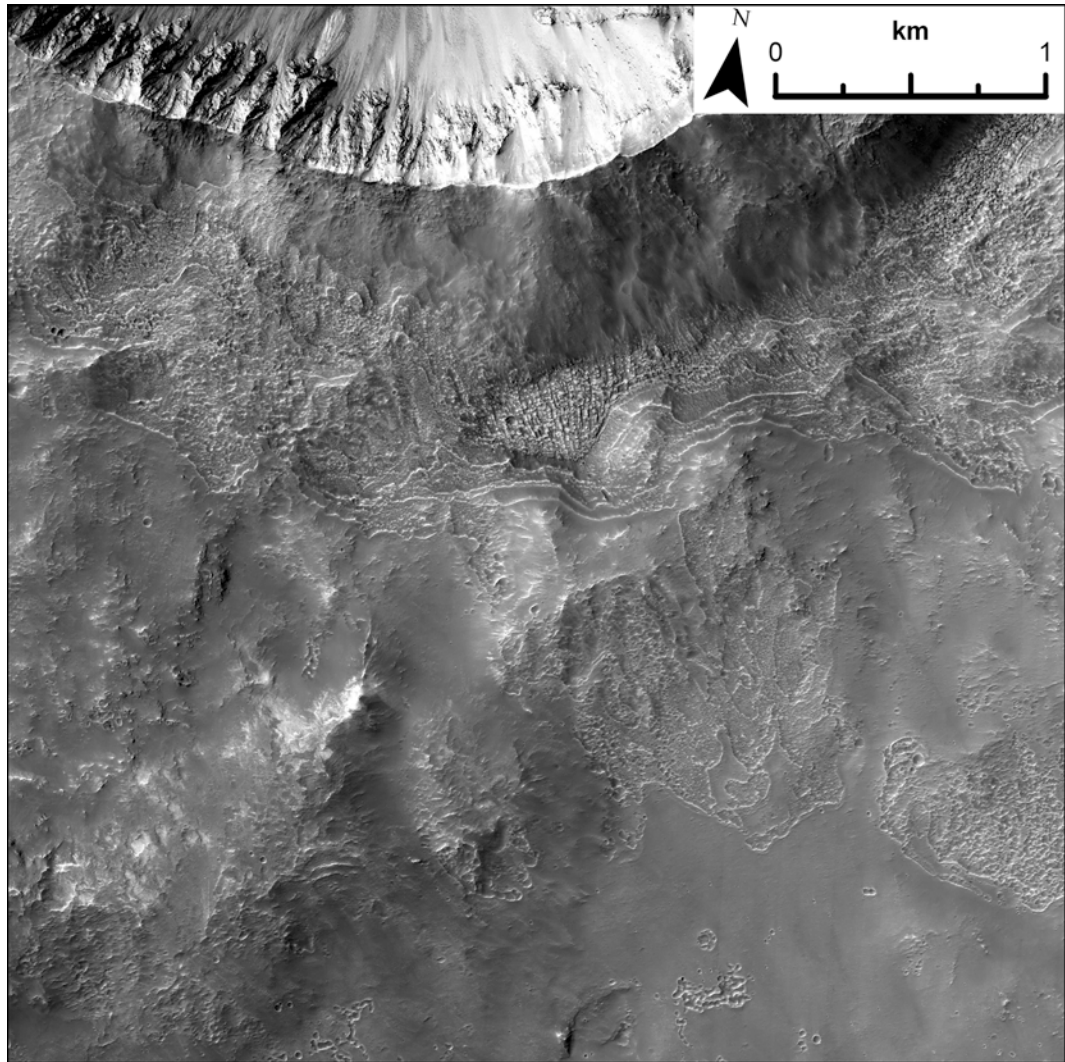


Fig. 12A: An unnamed 6.5-km diameter crater is found in Terra Sabaea (27.4°S, 59.1°E). Prominent geomorphic evidence of multiple layers of remnant latitude-dependent mantling (e.g., Schon et al., 2009a) is observed. Portion of HiRISE: ESP_014400_1525.

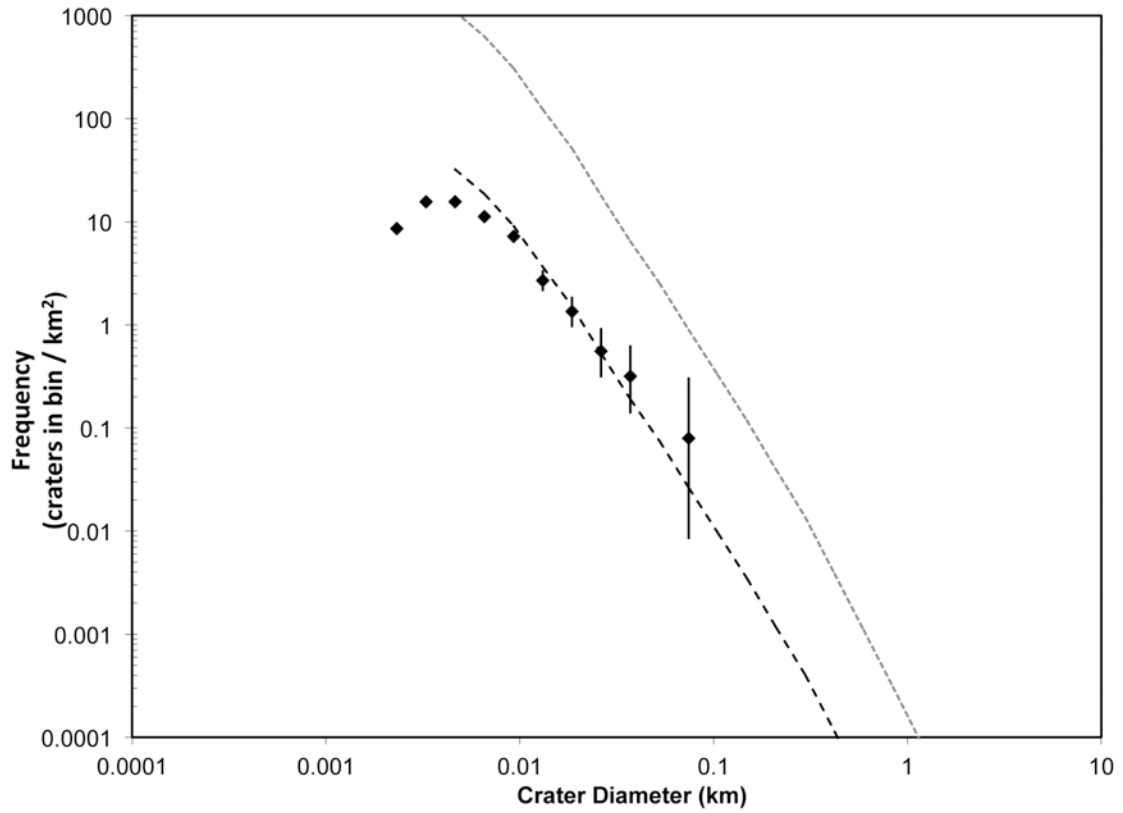


Fig. 12B: A crater count revealed 865 craters on 12.6 km² of smooth near-rim mantling material surrounding the unnamed crater (27.4°S, 59.1°E). Isochrons of Hartmann (2005) indicate a best-fit age of 7.9 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

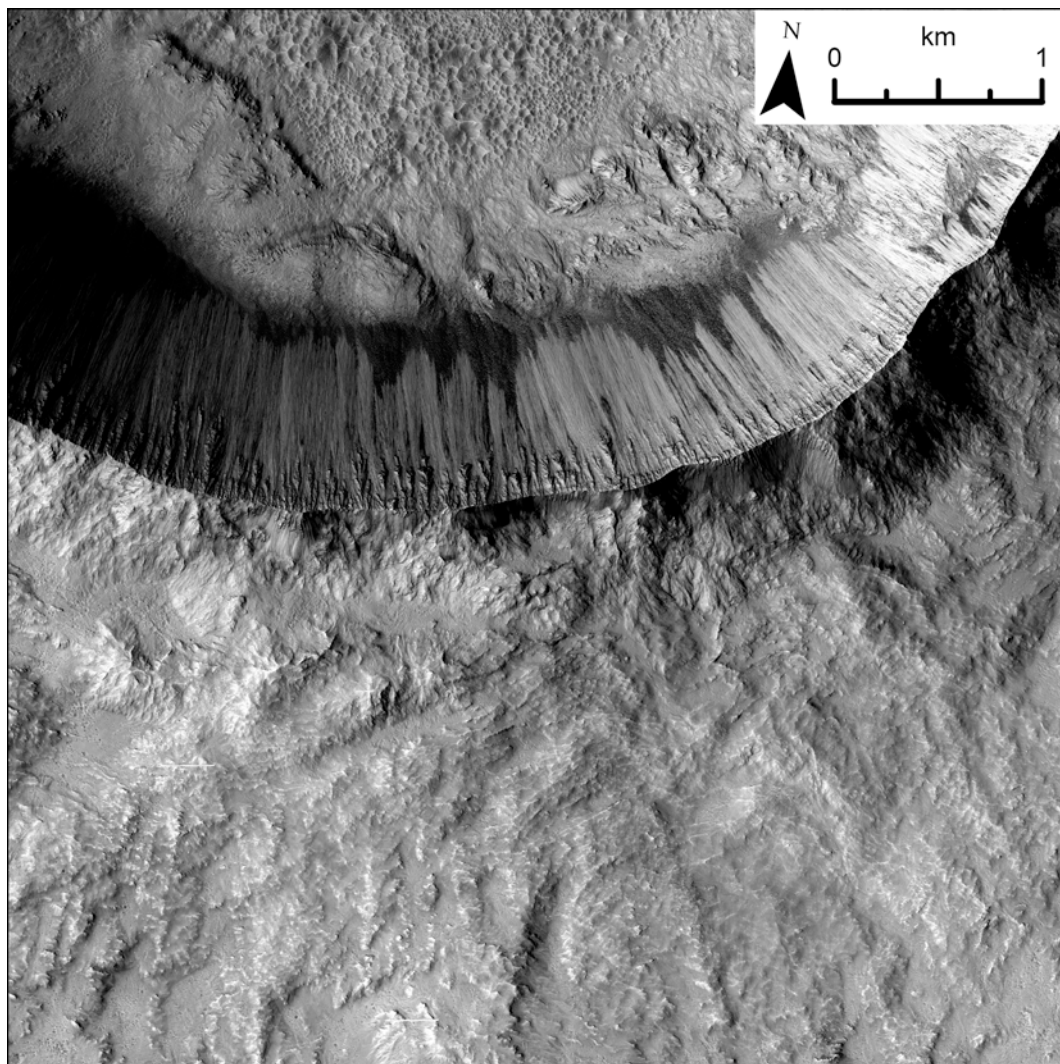


Fig. 13A: An unnamed 7.3-km diameter crater is found in Hesperia Planum (28°S, 108.9°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: PSP_004008_1520.

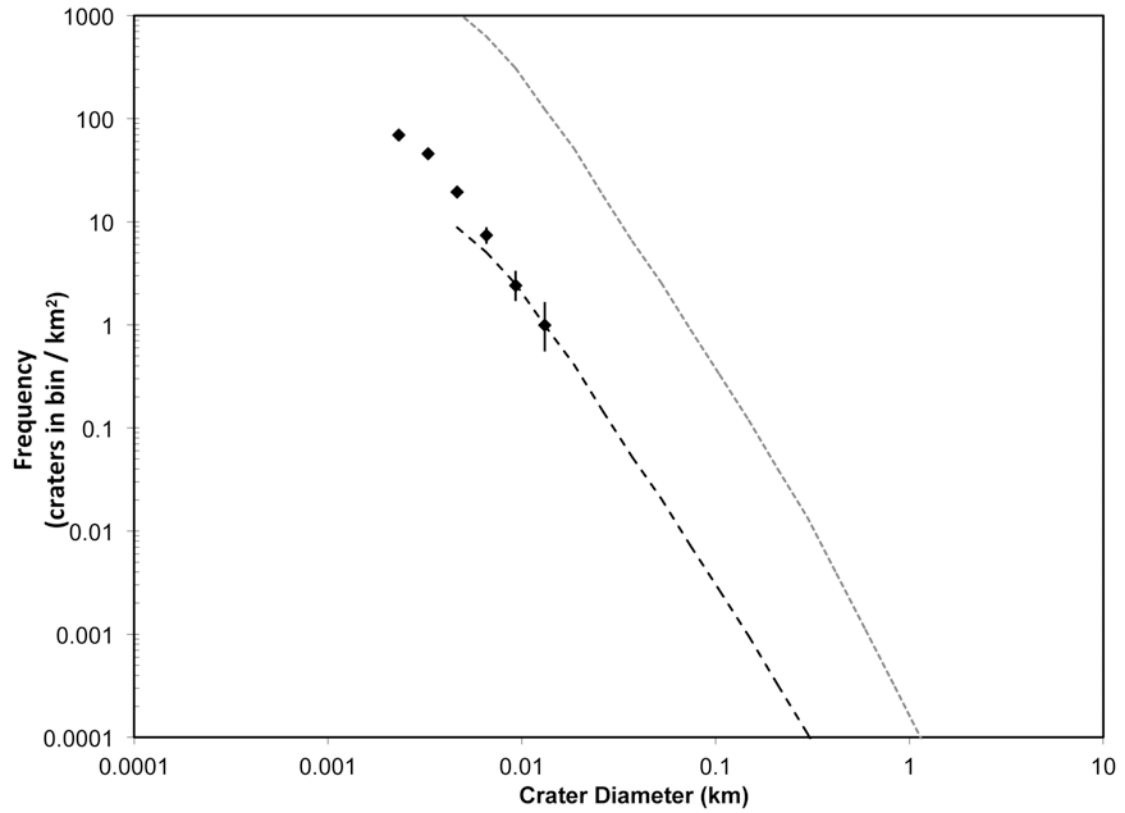


Fig. 13B: A crater count revealed 1,451 craters on 7.0 km² of near-rim deposits surrounding the unnamed crater (28°S, 108.9°E). Isochrons of Hartmann (2005) indicate a best-fit age of 2.2 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

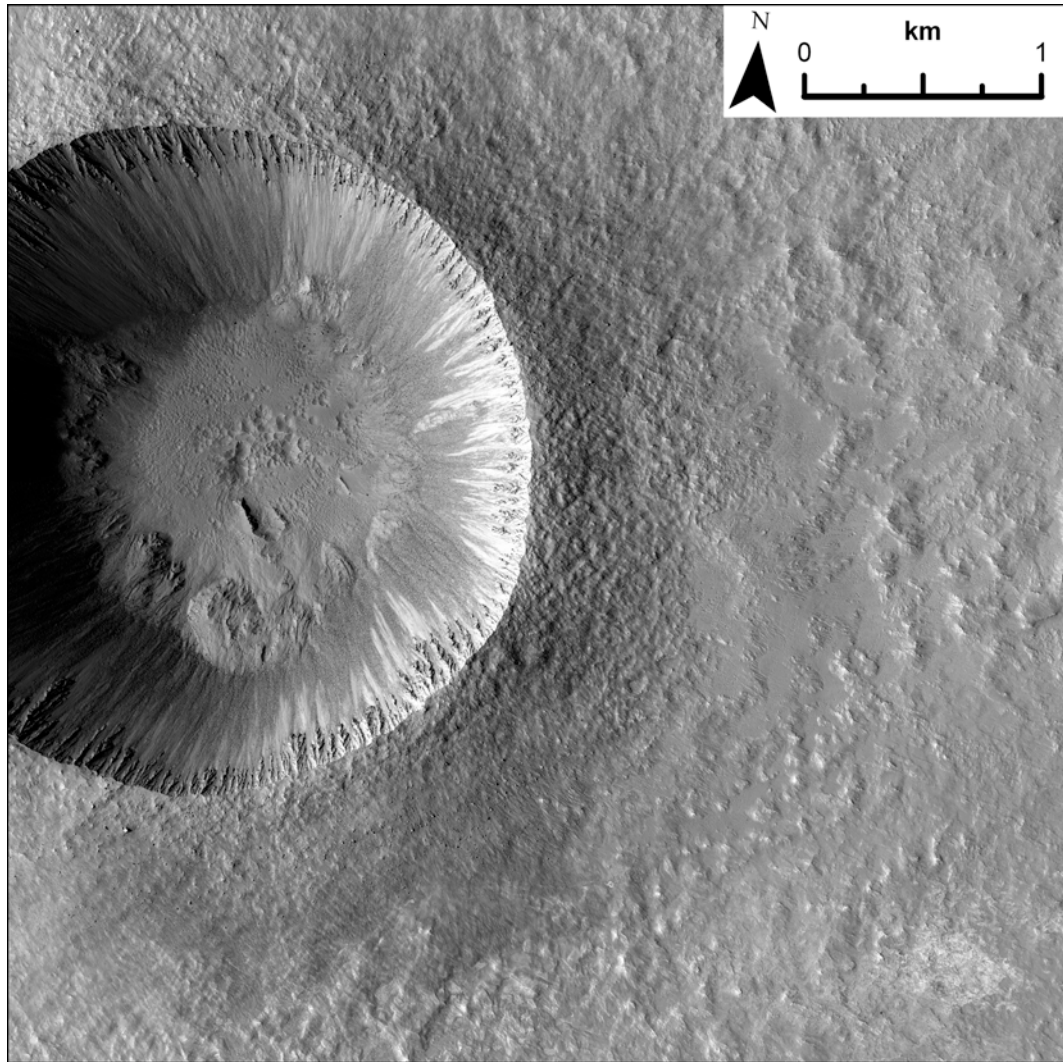


Fig. 14A: Zumba crater (28.7°S, 226.9°E) is a 2.9-km diameter rayed crater identified by Tornabene et al. (2006) in Daedalia Planum. No evidence of latitude-dependent mantling is observed. Zumba is a plausible launch crater for some martian meteorites. Portion of HiRISE: PSP_003608_1510.

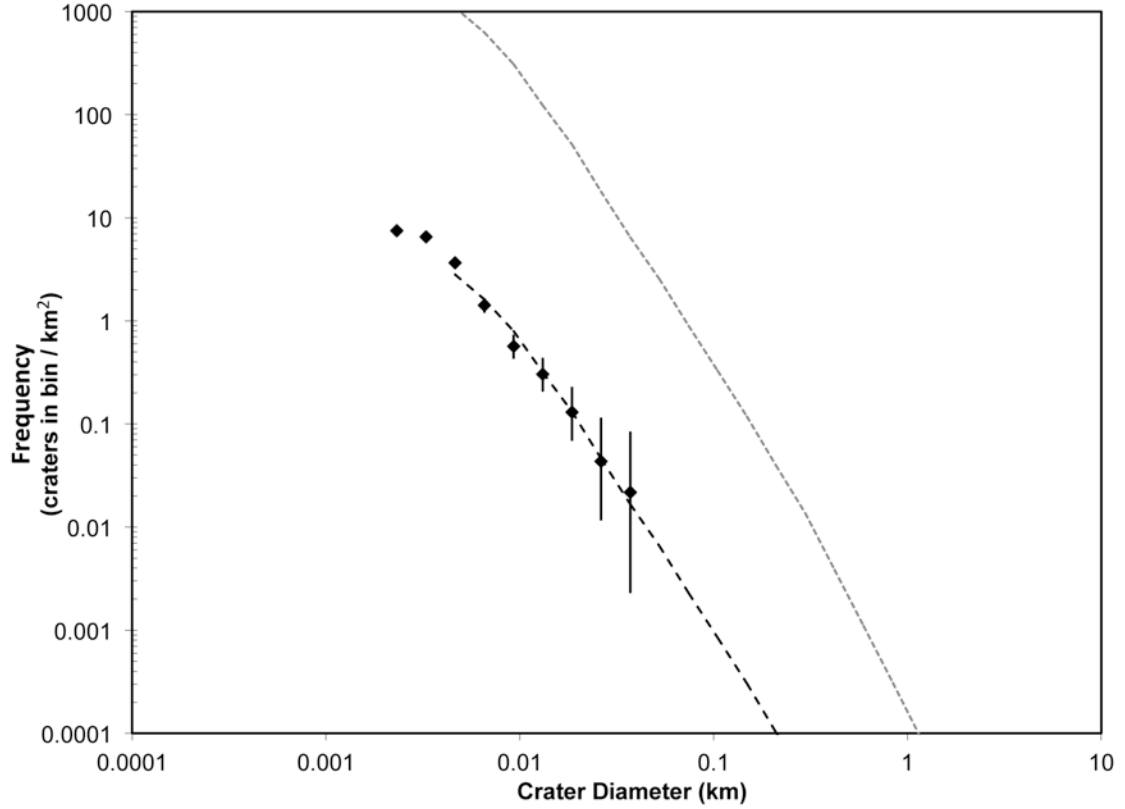


Fig. 14B: A crater count revealed 1,197 craters on 46.0 km² of near-rim deposits surrounding the unnamed crater (28.7°S, 226.9°E). Isochrons of Hartmann (2005) indicate a best-fit age of 0.7 Ma. Hartmann et al. (2010) report a crater retention age for Zumba crater of 0.1 to 0.8 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

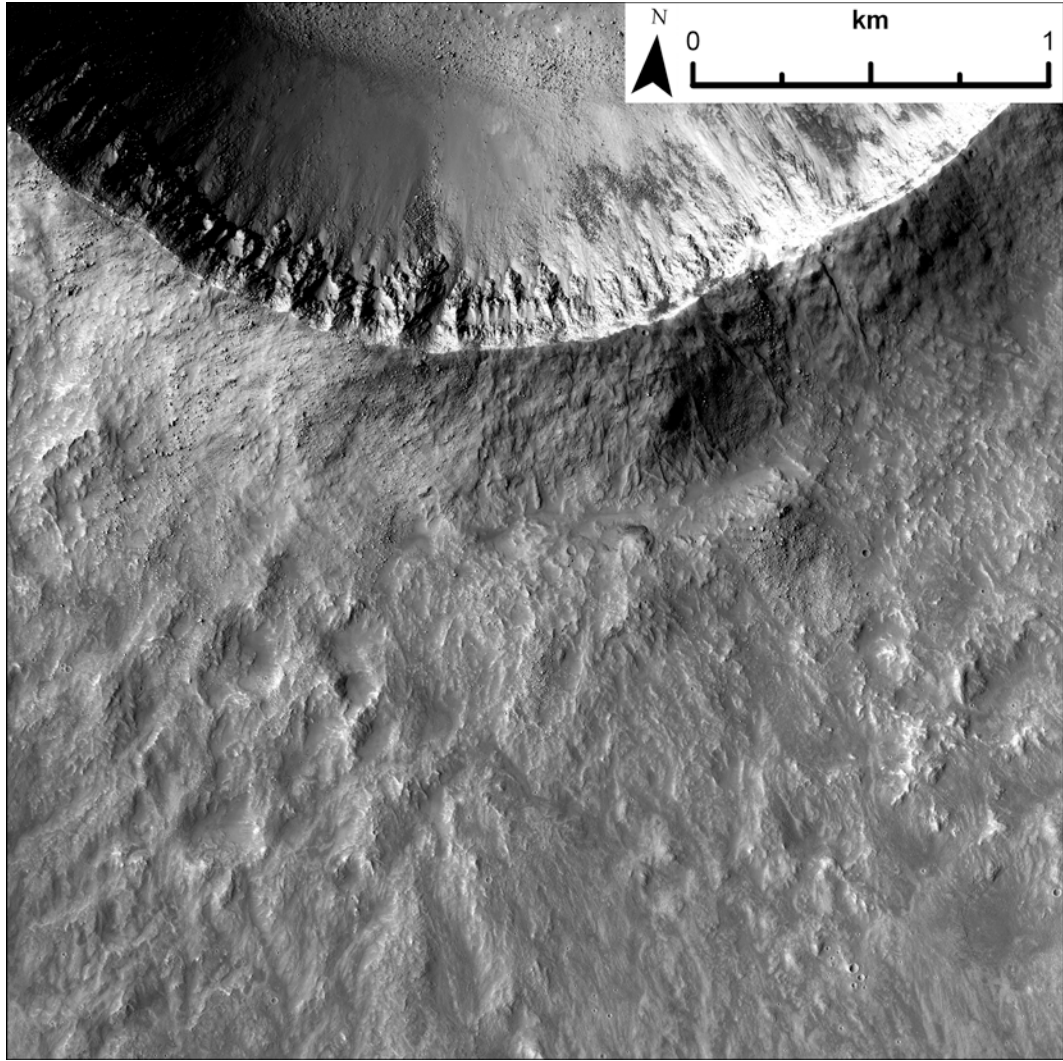


Fig. 15A: An unnamed 3.4-km diameter crater is found in Terra Cimmeria (29.5°S, 163.1°E). No evidence of latitude-dependent mantling is observed.

Portion of HiRISE: ESP_020752_1500.

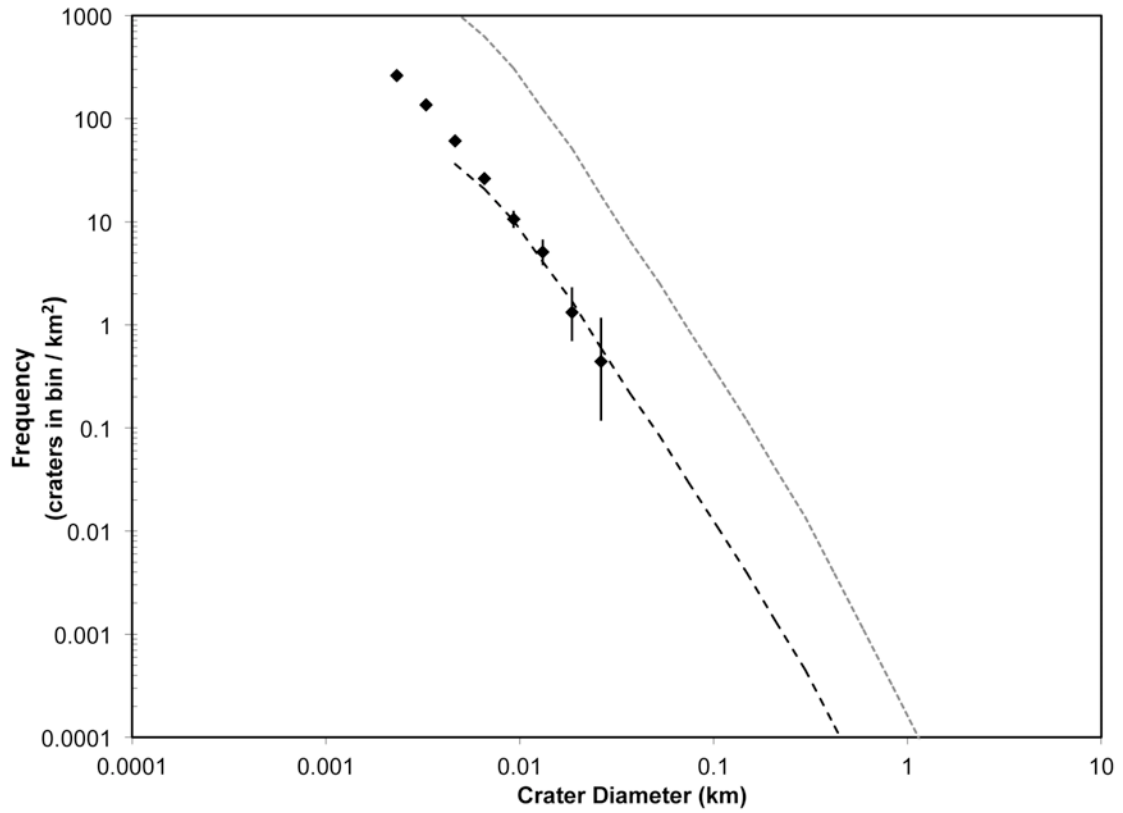


Fig. 15B: A crater count revealed 5,408 craters on 4.5 km² of near-rim deposits surrounding the unnamed crater (29.5°S, 163.1°E). Isochrons of Hartmann (2005) indicate a best-fit age of 0.9 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

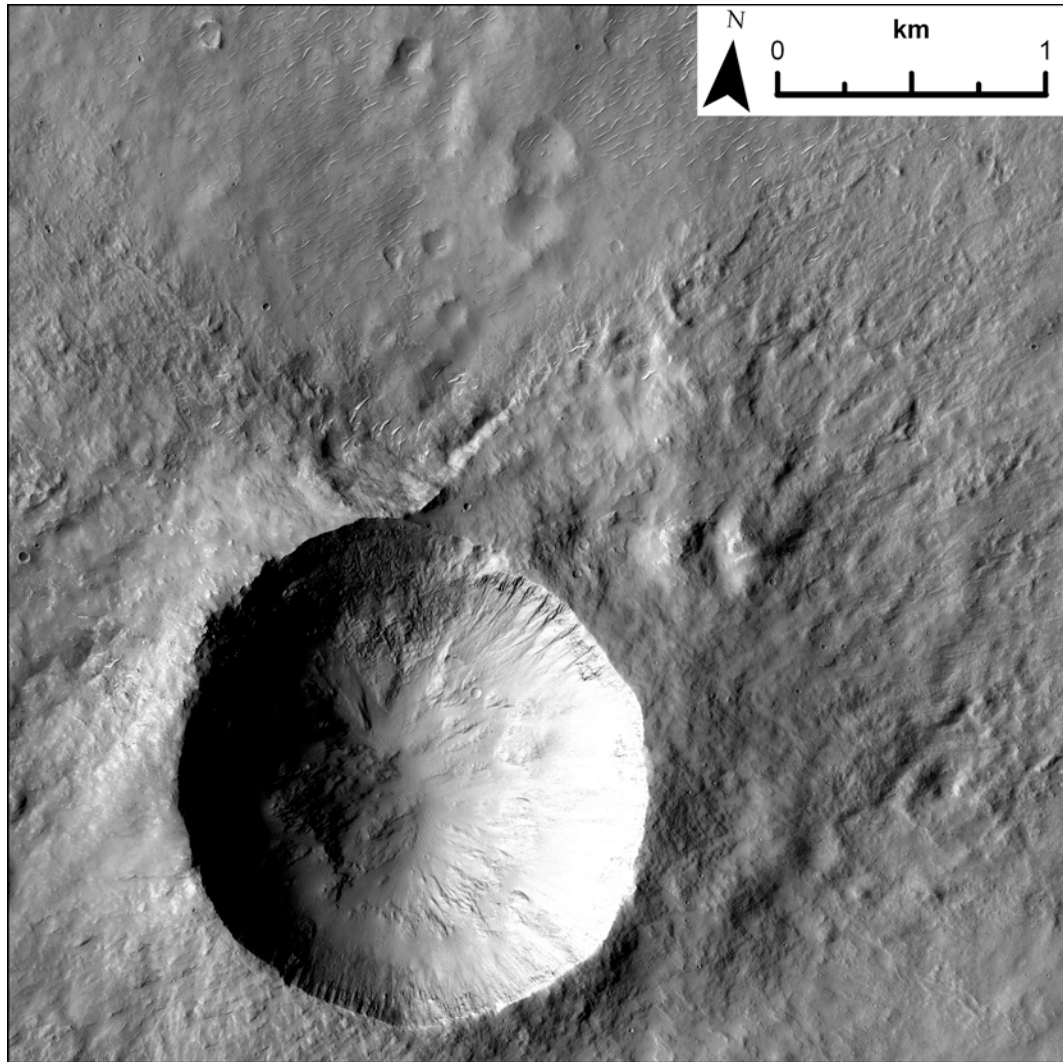


Fig. 16A: An unnamed 1.8-km diameter crater is found in Promethei Terra (32.2°S, 116.2°E) A triangular avoidance zone in the ejecta pattern (at top) indicates that this crater formed in an oblique impact. A small area of remnant latitude-dependent mantling is concentrated on the pole-facing crater wall in association with several gullies. Portion of HiRISE: ESP_020701_1475.

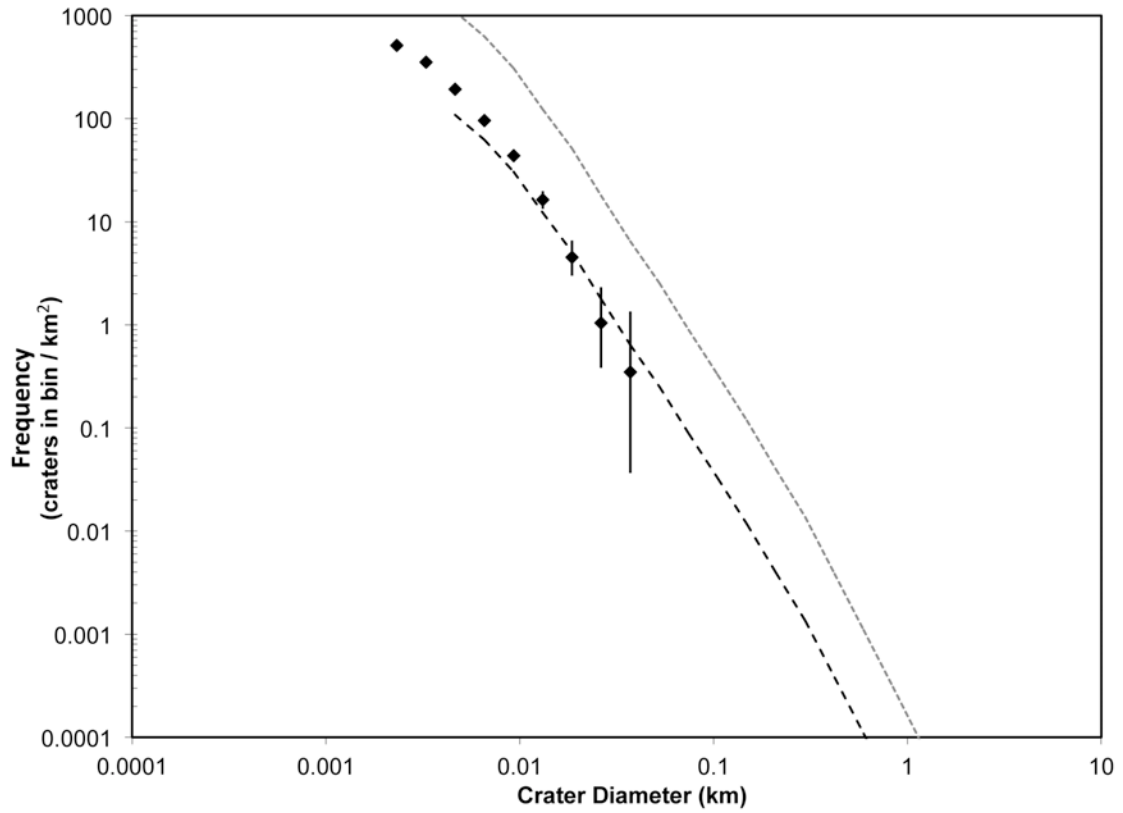


Fig. 16B: A crater count revealed 4,834 craters on 2.9 km² of near-rim deposits surrounding the unnamed crater (32.2°S, 116.2°E). Isochrons of Hartmann (2005) indicate a best-fit age of 26.8 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

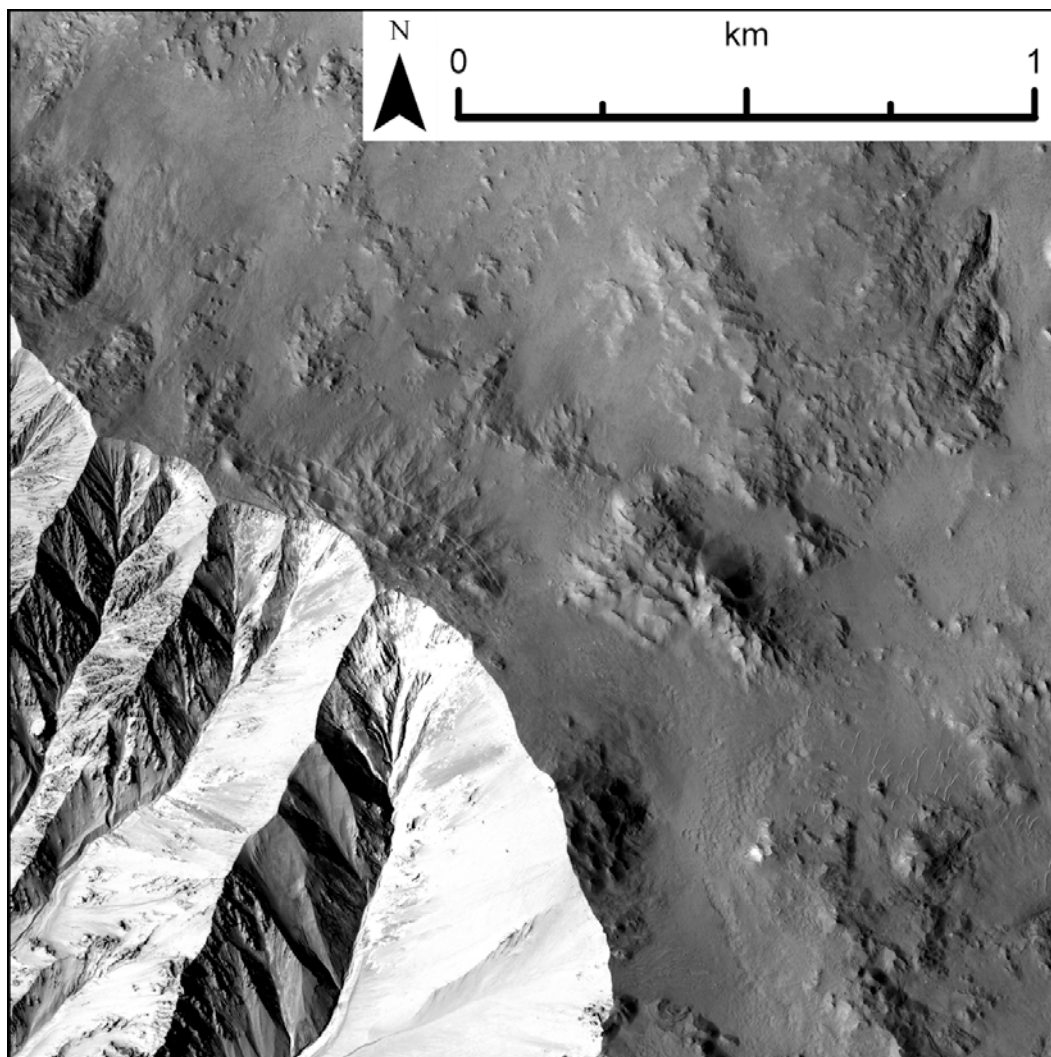


Fig. 17A: Gasa crater (35.7°S, 129.4°E) is a 7.0-km diameter rayed crater in Promethei Terra. Rays and secondaries from Gasa are observed on latitude-dependent mantling in the area (Schon et al., 2009b; Schon and Head, 2011b). Gully development within Gasa has been linked to impact into a debris-covered glacier (Schon and Head, 2011b). Portion of HiRISE: PSP_004060_1440.

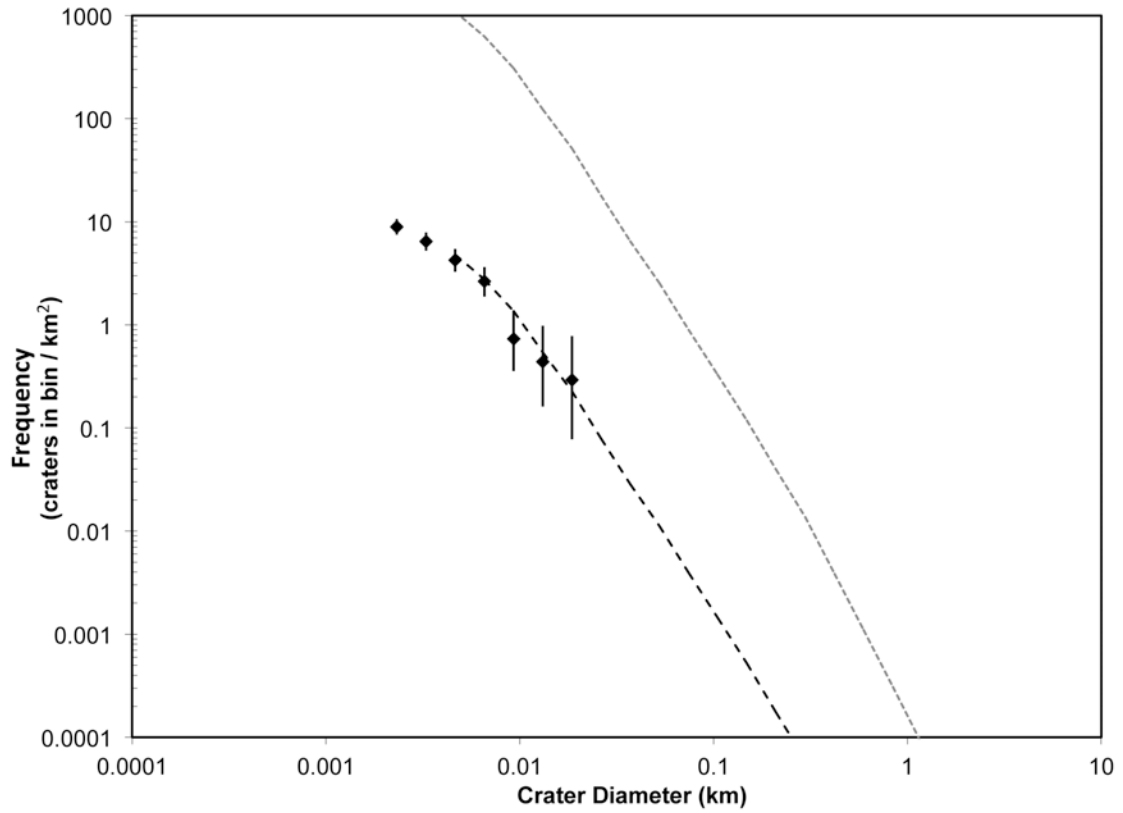


Fig. 17B: A crater count revealed 289 craters on 6.8 km² of smooth near-rim deposits surrounding Gasa crater (Schon et al., 2009b). Isochrons of Hartmann (2005) indicate a best-fit age of 1.2 Ma. The grey dashed line marks the Early Amazonian boundary of Hartmann (2005).

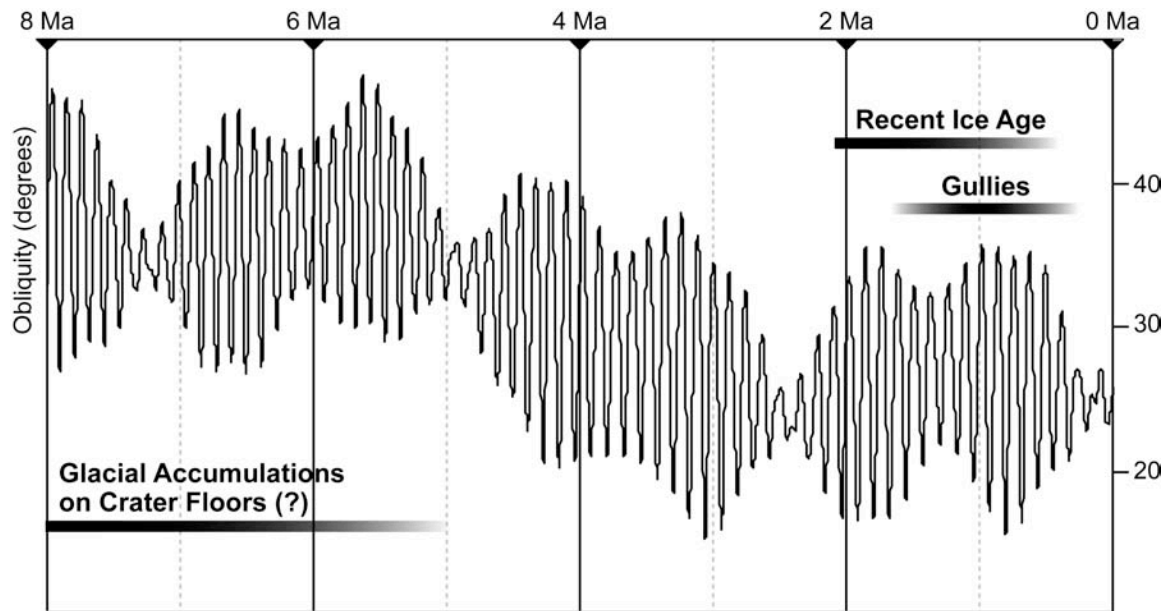


Fig. 18: Timeline of obliquity modulated (Laskar et al., 2004) latitude-dependent mantle deposition and modification during the Latest Amazonian period of Mars history. Mantling deposits are interpreted to have been emplaced in a latitude-dependent manner during an ice age coincident with the most recent period of enhanced obliquity (Head et al., 2003). Mid-latitude gullies, formed by the melting of ice-rich mantling deposits are coincident with the waning of this period (Head et al., 2008; Dickson and Head, 2009; Schon et al., 2009b). During the past 5 Myr obliquity has averaged $\sim 25^\circ$, but prior to approximately 5 Ma mean obliquity was $\sim 35^\circ$. Glacial accumulations on crater floors (e.g., Head et al., 2008; Dickson et al., 2011) are suggested to date from this period (Arfstrom and Hartmann, 2005; Berman et al., 2005; Berman et al., 2009). High latitude mantle terrain (e.g., Fig. 2; Fig. 3) with crater retention ages < 1 Ma correspond to the waning of the most recent ice age. Older more equatorial dissected latitude-dependent mantle terrain (e.g.,

Fig. 12) could correspond to the beginning of the most recent ice age, or alternatively to the transition to lower mean obliquity that occurred ~ 5 Ma. The chronological constraints (Fig. 19) support an equatorial ice source for deposition of the mantle (Levrard et al., 2004). Since high obliquity conditions were common in Mars' history, our observations of the current latitude-dependent mantle may represent only the most recent manifestation of a longer-term cyclic process that reconfigures surface ice reservoirs.

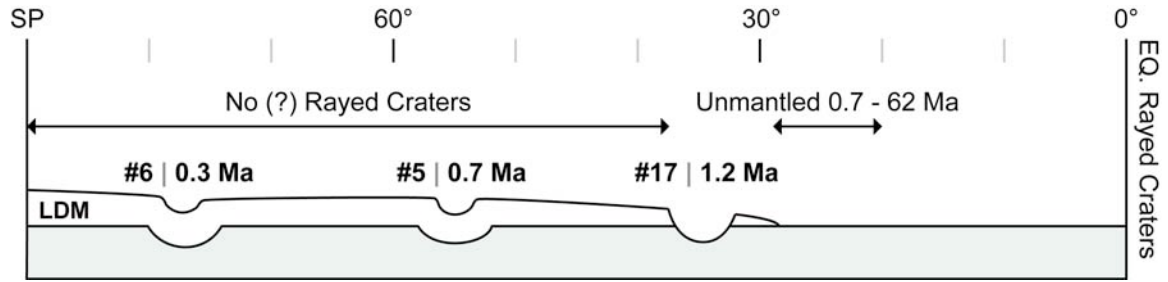


Fig. 19: This diagram shows the chronological constraints on the development of the latitude-dependent mantle derived from our observations. Two high-latitude craters (Fig. 5; Fig. 6) are draped by latitude-dependent mantling deposits. Crater counts on the polygonalized mantle surface at those locations reveal young crater retention ages. Gasa crater (Fig. 17) at 35.7°S superposes latitude-dependent mantling deposits which indicates that the mantle in this region is older than the high-latitude mantle. Un-mantled craters near the boundary of the LDM indicate that dissected mantle is restricted to the ~30-60° region (Fig. 1).

CHAPTER 7

PERSPECTIVES FROM DEPOSITIONAL ENVIRONMENTS: OUTSTANDING QUESTIONS AND EXPLORATION STRATEGY.

Introduction

You have reached the chapter typically known as the “synthesis chapter.” I prefer to think of these last pages as – what does it all mean? – and, where to from here? The preceding chapters have explored the elucidating sedimentological history of Mars; these specific vignettes provide some constraint on paleoclimate scenarios and allude to future exploration strategies.

Rather than recapitulating the conclusions of those studies, in this final chapter I endeavor to identify the unifying theme among the threads of Mars geologic history previously discussed, and propose outstanding scientific questions based on my thesis research and perspective. Finally, the chapter, and my thesis, closes with a short discussion of the strategic appropriateness of pursuing Mars sample return at this time.

In Chapter 1, Mars’ early Noachian geologic history was considered through the lens of lacustrine and deltaic deposits. In latter chapters, the focus shifted dramatically forward in geologic time to Amazonian Mars, where the development and evolution of gullies in the context of ice ages was

a central focus of investigation. The pervasive meters-to-tens-of-meters thick deposits of the ice-rich latitude-dependent mantle (Kreslavsky and Head, 2000; Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003; Milliken et al., 2003; Kostama et al., 2006; Schon et al., 2009) are a touchstone to Mars' icy present and recent past. These deposits also present the most visible, though arguably surficial, hallmark of the current configuration of Mars' ice reservoirs.

Since the Noachian, Mars' climate has evolved to the inhospitable cold polar desert environment of today: it exhibits remarkably slow erosion (Golombek and Bridges, 2000), extreme cold such that carbon dioxide seasonally freezes out of the atmosphere (Kelly et al., 2006), and very limited chemical weathering (Wyatt et al., 2004). The geomorphic signatures of this era, the longest in Mars history, are predominately the result of glacial processes.

The first identifications of glacial deposits were made with Viking data (e.g., Lucchitta, 1984; Squyres and Carr, 1986), but true detail was only revealed by extensive analysis with later datasets. As further data are gathered, we have to wonder – are some of these features static – the result of a progressive sequestration of ice from Mars' mobile cryosphere? Are gullies (e.g., Malin and Edgett, 2000) and Amazonian channels (Dickson et al., 2009) but the final vestiges of aqueous hydrology on Mars?

The Stratigraphic & Sedimentary Perspective

“I suppose it is tempting, if the only tool you have is a hammer,

to treat everything as if it were a nail.” – Abraham Maslow

The Psychology of Science: A Reconnaissance (1966), Ch. 2 p. 15

The spans of time and assortment of geomorphic processes covered in this thesis could be seen as disparate, from certain perspectives. What themes or ideas unite Mars’ early temperate Noachian (Phyllocian) with its volcanically-dominated, but outflow channel punctuated, Hesperian (Theiikian) – and finally with the long dry winter of the Amazonian (Siderikian)? The unifying scientific approach or method of the preceding chapters is stratigraphic in nature. Each of these eras, although they encapsulate enormously varied conditions, has geological problems that are amenable to stratigraphic investigation. In addition to stratigraphic methods, I would be remiss to not mention explicitly how powerful the paradigm of astronomically forced climate cycles (e.g., Ward, 1973; Hays et al., 1976; Touma and Wisdom, 1993; Head et al., 2003; Laskar et al., 2004) has been in my work.

In the preceding chapters, and in the work of others, the stratigraphic lens coupled with a keen focus on depositional environments has been fruitful in drawing new conclusions regarding longstanding research topics, and has

made other scientific problems tractable in new ways. However enamored I may be of this approach, it is important to remember and acknowledge that, as in most scientific endeavors, meaningful results are frequently the product of multiple ways of thinking coupled with disparate means of investigation. The present work and interpretations have benefited immensely from complementary advancements in, and discoveries enabled by, meter-and-sub-meter resolution imaging systems (e.g., Malin et al., 1992; McEwen et al., 2007a; McEwen et al., 2007b; Malin et al., 2010), mapping visible and near infrared spectrometers (e.g., Bibring et al., 2006; Mustard et al., 2008), gamma ray and neutron spectrometers (e.g., Boynton et al., 2002), laser altimeters (e.g., Kreslavsky and Head, 2000), shallow radars (e.g., Phillips et al., 2008), laboratory experiments (e.g., Chevrier et al., 2004), and computational models (e.g., Forget et al., 2006). Combining insights from all of these sources of data, including from ongoing research in planetary analogs (e.g., Marchant and Head, 2007; Fairén et al., 2010), will be necessary to fundamentally advance our understanding of Mars as a planetary system in the areas discussed below.

Outstanding Questions in Martian Geoscience

This section is meant to propose outstanding scientific research questions that have intrigued me during the course of my thesis research. While some questions are quite broad and perhaps equally applicable across

Mars' history, I have endeavored to separate questions related to Mars' earlier Noachian from later Amazonian time periods. The distinction is meant to reflect topics that may be related to the early temperate climate, and topics more apropos of the later hyper-arid and cold conditions dominated by ice-related processes. Some of these scientific questions are intentionally more specific in nature than the broad guiding questions found in community reports such as the planetary science decadal survey (Space Studies Board, 2011) or MEPAG reports (e.g., Hoehler and Westall, 2010). This list of questions, informed by my own perspective and interests, is by no means an exhaustive collection of the many queries that should be addressed by the scientific community in further research on these topics. For some of these questions that are more tractable I anticipate results from upcoming missions, while for others I provide background on the issue and discuss research challenges.

~Noachian questions:

- 1) *What was responsible for the termination of temperate conditions during the Noachian epoch?*

Data presented in Chapter 1 suggest that the lacustrine period in Jezero crater appears to have ended quickly. Many Noachian open-basin lakes have been identified in association with with valley networks (Fassett and Head, 2008b), and a global analysis of valley networks by Fassett and Head (2008a) indicates a synchronous termination for the

formation of these features (excepting special circumstances, such as formation on volcano flanks). The climate to form these features was probably the result of a thicker atmosphere and carbon dioxide greenhouse, but a sulfur-based warming mechanism (Halevy et al., 2007) has also been suggested. Recent detections of carbonate, both in situ (Boynton et al., 2009; Morris et al., 2010) and from orbit (Ehlmann et al., 2008) are an important advancement in understanding the early atmosphere, but at present the total volume of carbonates remains relatively unconstrained. The Mars Atmosphere and Volatile Evolution Mission (MAVEN), scheduled to launch in 2013, is designed to investigate a suite of atmospheric and ionospheric parameters and interactions (Jakosky, 2009). In particular, compositional and isotopic data from this mission will advance understanding of the overall evolution of Mars' atmosphere and climate and may allow greater insight into the abrupt termination of temperate conditions during the Noachian.

2) How much of Mars' original endowment of water has been lost via impact and solar wind erosion? At what rate and when did this occur?

This question is closely related to our overall understanding of the Noachian and potential habitability (e.g., Rampino and Caldeira, 1994; Fassett and Head, 2011). The early magnetic field would have contributed to retention of Mars atmosphere, but both the initiation and termination of the field are important to know in light of a denser solar wind early in

the history of the solar system. While large impacts may have temporarily created significant climate warming (Segura et al., 2002), basin-forming impacts are hypothesized to have stripped a significant portion of Mars' original endowment of volatile species (e.g., Melosh and Vickery, 1989; Ahrens, 1993). Results from the MAVEN mission can be expected to revolutionize our understanding of Mars' atmospheric evolution (e.g., Jakosky and Phillips, 2001).

3) *Were carbonates buried in a northern ocean, which could have led to a waning greenhouse and permanent collapse of an early “warm and wet” climate?*

Discussions regarding a hypothesized northern ocean have been occurring for some time (e.g., Parker et al., 1989; Baker et al., 1991; Malin and Edgett, 1999; Clifford and Parker, 2001; Carr and Head, 2003). Putative shorelines (e.g., Parker et al., 1989) and the position of sedimentary deposits (Di Achille and Hynek, 2010) have been interpreted as supporting such an ocean, and present topography is potentially consistent with some of the suggested shorelines (Head et al., 1998; Head et al., 1999). Supposing a northern ocean occurred and persisted, could it have sequestered enough carbon to change the climate? Since nearly all of these carbonate deposits would be overlain by younger deposits, including Hesperian volcanic plains and the Vastitas Borealis Formation, this will be challenging to evaluate volumetrically. Alternatively, any ocean that

did exist may have been acidic and not have precipitated carbonates (Fairén et al., 2004).

4) *How variable were temperate Noachian conditions? What does this imply regarding astrobiological potential?*

Astrobiology is an undeniable motivation for Mars exploration, both in the popular imagination and in the research community (NRC, 2007). Results presented in chapter 1 indicate that surface flow and climate stability were required to construct a progradational delta, but the residual accommodation space suggests this period was limited. The duration of these potentially “globally” habitable conditions is of keen interest to astrobiologists because later, and perhaps more enduring, habitable environments are likely to have been niche environments akin to terrestrial extremophile environments. In order to put this question in context, additional terrestrial research is needed to characterize early Earth conditions and develop additional insights from the earliest fossil record (e.g., Schopf, 1999; Hazen, 2005).

5) *How pervasive and enduring were subsurface hydrothermal systems following the end of “hospitable” surface conditions?*

Thermal anomalies on the surface resulting from volcanic processes have not been observed in the instrumental record – however, geologically young volcanic flows have been identified (Hartmann, 2005). Given present surface climate and radiation conditions, it is likely that any

environments on Mars that are presently habitable would be in the subsurface and associated with a geothermal heat source capable of supporting a chemosynthetic metabolic pathway (e.g., Boston et al., 1992). The duration, distribution, and interconnectedness of such environments are unknown; however, their existence is supported by occurrences of low-temperature, aqueous alteration products in the rock record (e.g., Poluet et al., 2005).

~Amazonian Questions:

1) Can atmospheric methane be verified by in situ investigation?

Though earlier detection claims had been made on the basis of OMEGA data (Formisano et al., 2004), in 2009 Mumma et al. presented the first persuasive evidence of atmospheric methane on Mars using observations made with infrared spectrometers on ground-based telescopes. However, the detection methodology of Mumma et al. (2009) has been disputed by Zahnle et al. (2011), who suggest that terrestrial methane absorption features make ground-based observations unreliable. The Mars Science Laboratory's Sample Analysis at Mars (SAM) toolkit for analytic chemistry will enable in situ observations of atmospheric methane presence, abundance, and isotopic ratios (Webster and Mahaffy, 2011). Following MSL in 2011 and MAVEN in 2013, the NASA-ESA Trace Gas Orbiter mission, with a nominal launch in 2016, will conduct a

campaign of atmospheric observations and measurements to delineate the trace gas constituents of the martian atmosphere and any seasonal variations (e.g., Zurek et al., 2011).

- 2) *If methane is currently present in the atmosphere, what source, or multiple sources, is responsible for its production and release? Additionally, what processes influence its concentration and destruction?*

Assuming in situ confirmation of atmospheric methane, the question of its provenance will become central. Possible sources may include active volcanism, the geological process of serpentinization, subsurface microbial life, and ancient methane accumulations of either biological or geological origin. Future missions (MSL and the Trace Gas Orbiter) may constrain the geographic distribution of sources, which would prove useful in discriminating between methane source hypotheses. Similarly, time-series observations from these missions will be important to understand the life cycle of atmospheric methane since reported observations (Mumma et al., 2009) are inconsistent with the expected photochemical lifetime of atmospheric methane – implying perhaps a powerful, unknown destruction mechanism (Lefèvre and Forget, 2009).

- 3) *What are the characteristics, volume, age, and history of mid-latitude ice deposits?*

While decades of research have identified and investigated non-polar glacial features (e.g., Head et al., 2005; Head et al., 2010), current

understanding of the development and geologic history of these features remains incomplete. The likelihood of cold-based glaciation, which leaves little geomorphic signature, makes conclusive identification of expired glaciations of this sort quite difficult. Additionally, chronologies based on relative crater densities have their largest uncertainties for intermediate ages. These twin challenges make it difficult to reconstruct the complete glacial history of Mars. However, recent mapping of drop moraines (e.g., Head and Marchant, 2003; Garvin et al., 2006; Shean et al., 2007), debris covered glaciers (e.g., Head et al., 2010), volcano-ice interactions (Head and Wilson, 2002; Head and Wilson, 2007), pedestal craters (Kadish et al., 2008; Kadish et al., 2010; Schon and Head, 2011), and layers of latitude-dependent mantling deposits (Schon et al., 2009) all suggest ice deposits variable in their initial deposition, maximum volume, age, and extent. It is difficult to discern to what degree the presently observable glacio-geomorphic features reflect only preservation potential rather than the complete original suite of glacial systems. Similarly, while evidence for individual advances and retreats is abundant, it is unknown to what degree this record and the implied chronology is complete. Comprehensive meters-scale imagery, improved radar instruments (cf., Holt et al., 2008), and extensive geologic mapping studies can be expected to further improve our understanding of these deposits, but known unknowns relating to the precise chronology (likely driven by un-constrained orbital

forcing) and unknown unknowns (related to the possibilities of cold-based glaciation) will almost certainly remain.

4) *Has the Amazonian been characterized by a progressive sequestration of ice volume in mid-latitude reservoirs?*

This question is, in part, a conjugate of the previous question.

Setting aside the precise step-wise history of Amazonian glaciations – during these events, has there been a long-term removal of ice from the dynamic surface environment to debris-covered and subsurface environments that respond only weakly to surface conditions and climate in the present era? Observations with radar have shown lobate debris aprons to be debris-covered glaciers (Holt et al., 2008; Plaut et al., 2009). Geological evidence suggests that these features predate, perhaps substantially, the recent ice age (e.g., Morgan et al., 2009). Therefore, these features have survived significant orbital/climatic variations (Laskar et al., 2004) and could represent a long-term repository for ice. Tallying the volume, age, and long-term survivability of these deposits will reveal important details regarding the evolution of Mars' cryosphere.

5) *Do gullies meet an extremophile-based definition of a habitable environment?*

Mid-latitude gullies have been classified as a special region (Beaty et al., 2006), which would require more rigorous planetary protection measures for an in situ investigation. Based on multiple lines of evidence, a scientific consensus supports that their formation is dependent on the

presence of aqueous fluids and occurrences of fluvial sediment transport. Top-down melting of surficial ice and snow is a primary source of meltwater for gullies. The youthfulness of these features and their direct relationship with recent occurrences of liquid water makes them compelling targets for astrobiological investigation. Yet, is it correct to consider gullies as potentially habitable environments? Terrestrially there are many cryptobiotic species, some quite complex, that can survive interludes of extremely harsh environmental conditions. For example, species of phylum Tardigrada are particularly known for their ability to survive extreme desiccation, as well as temperature, radiation, and pressure variations (Nelson, 2002). However, the duration and recurrence interval of environmental conditions amenable to even the hardiest of extremophiles are likely to be far more challenging in gullies on Mars than in even the harshest terrestrial environments. More research will be needed on gully microenvironments (e.g., Marchant and Head, 2007), as well as the limits of anhydrobiosis (e.g., Crowe et al., 1992), in order to move towards any conclusions to this question.

Is Mars Sample Return the Right Goal, Right Now?

Sample return has been a holy grail of planetary science since the 1970's. Recently, sample return from Mars was codified as the highest priority goal by the National Research Council's decadal survey for planetary

science –echoing a call from the influential Mars Exploration Program Analysis Group. Is now the right time for a risky and expensive sample return mission, as they are suggesting? Given that a robust and fast-paced Mars exploration program is required to maintain the tempo of new discoveries and continued American leadership in planetary exploration, recent mission successes and the current fiscal situation suggest to me that a series of smaller missions, both orbiters and rovers, would be a better strategy.

Over the previous five decades, nearly two score of spacecraft have been launched with the goal of exploring Mars (Garber, 2009). While early efforts were beset by repeated failures, at least a dozen more recent missions must be hailed as staggering successes – from the early pioneering work of the Viking orbiters and landers, to the recent longevity and endurance of the Mars Exploration Rovers. These missions have revealed surface features, chemical compositions, rock mineralogies, global topography, and other important measures of the planet.

Like the pioneering expedition of Lewis and Clark that laid the groundwork for westward expansion, these missions have suggested new avenues for exploration and discovered environments that were potentially once habitable. These data have been used in extensive landing site selection studies in preparation for the nearly SUV-sized Curiosity rover that is scheduled to launch this year (Grotzinger, 2009; Grant et al., 2010; Golombek

et al., 2011). With any luck, Curiosity will spend several years characterizing the geologic history and potential habitability of its field site, Gale crater. It has the ability to address many pressing questions in planetary science: Were sediments transported and deposited by flowing streams? Was there a long-lived hydrothermal system underground? How and when did environmental conditions change to the cold, arid Mars of today?

It is true that returned samples are immensely valuable for science. We know this from the Apollo samples that continue to teach us so much about the Moon (e.g., Saal et al., 2008; Hauri et al., 2011), as well as from the first rocks dredged from the Earth's mid-ocean ridges. Samples in terrestrial laboratories can be subjected to experiments un-thought of, using tools not-yet developed, when the materials were originally collected. Multiple laboratories can verify results and collaborate. Obviously, much larger apparatuses can be brought to bear in analyses on Earth than can be crammed into a rover.

However, the multi-billion dollar costs and multi-decade timescale of the proposed Mars sample return mission are staggering, and come at a very real scientific price of forgone alternative opportunities. The sequence of three component missions proposed to accomplish sample return in the new decadal study is not scheduled to return samples until sometime beyond the 2013-2022 decade. New technologies, such as for the Mars ascent vehicle, will be required. The study committee also assumed a deep and long-term

partnership with the European Space Agency (i.e., Mars Exploration Joint Initiative), which could be difficult to coordinate and sustain over the course of such an endeavor, especially given the current budgetary constraints and uncertainties on both sides of the Atlantic (Altman and Haass, 2010; Erlanger, 2011).

While the decadal survey planned for a rising planetary science budget, the Administration's proposed FY12 budget calls for sustained cuts in NASA's planetary science division (Green, 2011). Given these realities, embarking on a Mars sample return campaign now is simply unaffordable and incompatible with a balanced program of planetary exploration. The past decade's success at Mars is the result of a flotilla of targeted missions following up on the discoveries of previous spacecraft, and the continuation of this strategy remains the best way forward for the Mars exploration program.

Sample return remains a worthy long-term goal and should be supported through affordable and sustained investment in advanced spacecraft technology (e.g., Hamilton et al., 2011). Only a fiscally responsible and sustainable program of sample return missions will be capable of delivering to terrestrial laboratories the multiple suites of samples required to accomplish the scientific goals of sample return regarding Mars' planetary evolution, habitable environments, and climate over geologic time.

Development of enabling technologies should be invigorated, of course. However, the tempo and success of NASA's Mars exploration program should

not be sacrificed by going all-in on an unaffordable multi-decade sample return mission right now.

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